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ROCK GLACIERS IN THE COLORADO FRONT RANGE

BY RONALD L. IVES

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ABSTRACT

Physiographic studies in the Colorado Front Range area disclose a number of rock glaciers in various stages of evolution. Detailed investigations show that the theories of Capps are substantially correct for rock glaciers originating in abandoned cirques, but that a number of factors not considered by him may be of importance. These factors are briefly discussed. Reconnaissances in other mountain areas suggest that rock glaciers and similar features are very common wherever relatively recent alpine glaciation has taken place.

INTRODUCTION

GENERAL STATEMENT

Discovery, in 1932, that the terminal moraine of Fair Glacier, at the head of Arapaho Creek, Grand County, Colorado, had a slow motion independent of that of the glacial ice, led to a search for other detrital accumulations having similar motions. This search, carried on in conjunction with studies of alpine glaciation, has resulted in the discovery of many small, but typical, rock glaciers, and of a still larger number of marginal cases, believed to be ancient rock glaciers, now nearly or entirely static.

LOCATION AND ACCESSIBILITY

The area studied in detail is located near the Continental Divide in Boulder, Grand, Jackson and Larimer counties, Colorado, at the headwaters of the Platte, Colorado, Laramie, and Cache-La-Poudre rivers. Topography is shown on the United States Geological Survey Rocky Mountain National Park quadrangle. Although the rock glaciers in this area are near timberline, several miles from, and several thousand feet above, the nearest automobile road, they may be reached by trail in a few hours from the nearest road point. Local inquiry is advisable, for the trails are currently (1940) being rebuilt and rerouted.

PHYSIOGRAPHIC SUMMARY

All of the rock glaciers here described are found within glaciated valleys cut into the granite, schist, and gneiss of the Front Range during Wisconsin time. In a few areas, the crest of the range is veneered with vol-

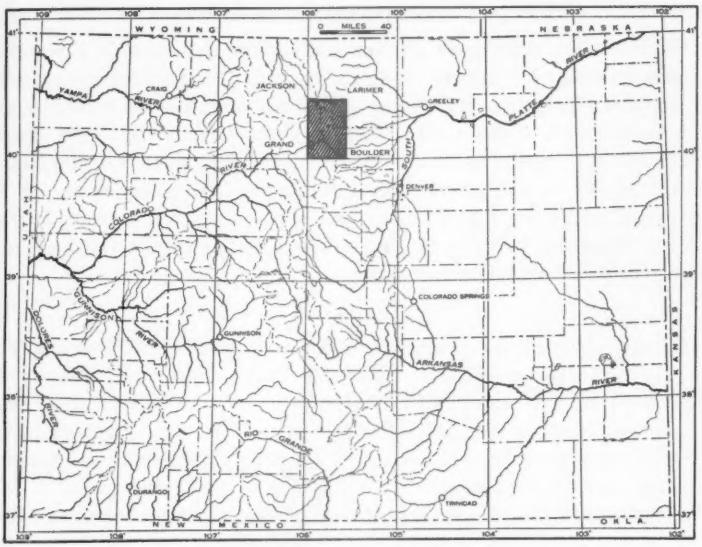

FIGURE 1.—*Index map*
Rock glaciers here described are within the shaded area.

canics, probably of early and middle Tertiary age, but in every instance the bottoms of the valleys are cut into the Front Range pre-Cambrian complex, which has been adequately described by Bastin and Hill (1917). In no instance has a rock glacier been found in a valley not glaciated in Monarch (latest Wisconsin, Ives, 1938a) time; those best developed are found in cirques last filled with ice during the Neva glacial stage, roughly contemporaneous with the "Little Ice Age" recently reported by Matthes (1939).

ACKNOWLEDGMENTS

Field work in this area was made possible by two grants from the Penrose Fund of the American Philosophical Society, and by a grant from the University of Colorado Museum, which also furnished materials and laboratory facilities. The writer is grateful to Dr. P. G. Worcester, of the University of Colorado, and to Dr. S. R. Capps, of the U. S. Geological

Survey, for many helpful discussions of the field evidence. Friendly assistance, given without hope of reward by many residents of the mountain region, greatly expedited the field work.

ROCK GLACIERS

GENERAL STATEMENT

Studies of the many rock glaciers and similar structures in this region suggest a regular life history for them, beginning with a runaway moraine or unstable talus slope, and ending with a stable and static detrital mass. The early stages of this evolution, from youth to maturity, were clearly described by Capps (1910) in his report on the rock glaciers of the Nizina region, Alaska. Descriptions of the later stages, from maturity to senility, are based on evidence to be presented here. The rock glaciers in Fair Valley, Hitchens' Gulch, and Hague Creek Valley typify youthful, mature, and senile stages, respectively.

FAIR VALLEY

General description.—The Fair Valley rock glacier (SE¼ sec. 3, T. 1 N., R. 74 W., Rocky Mountain National Park quadrangle), actually the runaway terminal moraine of Fair Glacier,[1] represents the youthful stage in the evolution of a rock glacier occupying a cirque. Located in the Crater Lake Hellhole, at the head of Arapaho Creek, in Grand County, this structure is approximately 1000 feet long, 650 feet wide, and more than 150 feet thick at the downstream end. Estimates based on extrapolated valley profiles indicate that this thickness does not exceed 500 feet, but is more than 200 feet. The average slope of the upper surface of this rock glacier is about 20 degrees; that of the nose, where it forms the south shore of Triangle Lake, is about 45 degrees, and exceeds the average angle of rest of nearby talus by about 10 degrees. Relations of glacial ice, rock glacier, and lake are shown in Plate 1, an oblique photograph taken from the easternmost gendarme at the south end of the Lindbergh Peak "razor edge". A longitudinal section of this structure is shown in Figure 2.

Component fragments.—Fragments composing the Fair Valley rock glacier range in size from masses 25 feet in diameter (of which there are several score) to particles so small that they will remain suspended in water for many hours. On the surface, and to a depth of more than 10 feet below it, near the nose of the structure, rock fragments average more than 2 feet in diameter. Upslope, near the terminal margin of Fair Glacier, the average size of the fragments is smaller. In this rock glacier,

[1] Neither Fair Glacier nor Triangle Lake, at its foot, are shown on this map.

and in all others studied in this area, the average size of the components seems to decrease rather rapidly with depth.

Interstitial conditions.—Between the components, in the lower part of the structure, are large open spaces, into which a man can crawl. These

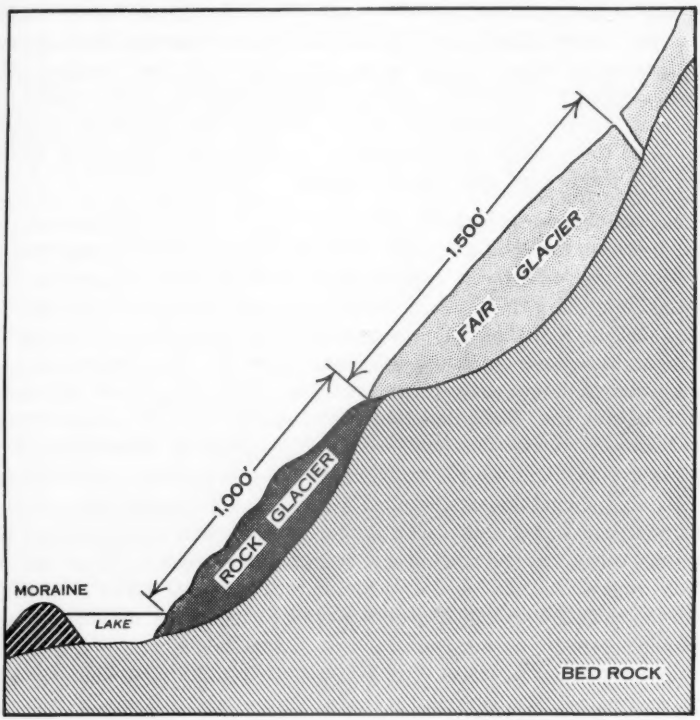

FIGURE 2.—*Longitudinal section of Fair Valley rock glacier*

openings decrease in size with depth below the surface, and with distance upslope. In the upper part of the rock glacier, the interstices are filled with compacted rock flour. Although its component particles remain suspended in water for hours, this material settles to a hard cementlike mass as resistant to erosion and deformation as freshly-exposed Pierre shale. Where this filling is present, there are no ridges on the surface of the rock glacier, suggesting that a mud matrix, of this type at least, inhibits, rather than assists, rock glacier creep. This ridgeless area is clearly visible in Plate I, just below the banded ice of Fair Glacier.

Very little interstitial ice is present here late in a normal summer, even where rather large snowbanks rest atop the rock glacier for a large part of the year. This is in direct contrast to conditions observed by Capps (1910) in Alaska, but in general agreement with those observed elsewhere in this area, and in other parts of the Southern Rocky Mountains.

Many of the interstices contain ponded water, on the surface of which ice forms almost nightly, even during the warmest part of the summer. The bottoms of many of these ponds are covered with fine compact silt. Observations over a number of years suggest that the water content of this rock glacier, as indicated by the number and estimated average volume of these ponds, is substantially constant, even though the water level in a single pond may change greatly.

Alimentation.—Materials forming the Fair Valley rock glacier have two known sources, and two other probable sources, of minor importance. A considerable part of the detritus is morainal material, released by the melting front of Fair Glacier. A rather recent recession of this glacier (Ives, 1938a, p. 1063-1064) has undoubtedly released much material, formerly ice-bound, to the rock glacier below it. A continual rain of rocks from the steep granitic walls of Fair Valley is also an important source of supply. The type area for such a rain of rocks is the Chasm, just below the precipitous East Face of Longs Peak, in Rocky Mountain National Park. During the early mountain spring (June), when snows are melting, mountaineers climbing in this area find crash helmets desirable items of equipment. Similar phenomena have been reported by Tyrrell (1910), and are well known to alpinists everywhere. Rough calculations suggest that as much as a ton of rock may fall into the drainage of the Fair Valley rock glacier daily during the summer. Although small rock slides are by no means infrequent in the higher parts of the Front Range, no good evidence of them is present here. As rock slide deposits are removed within a very few years by both ice and rock glacier motion, absence of such evidence indicates only that no important slides have occurred here recently. Direct observations by the writer, frequent since midsummer of 1932, produced no evidence of slides. Studies of an extensive series of photographs, now in the University of Colorado Museum, taken during the progress of various high-country studies by R. D. George, Ed. Tangen, the late Junius Henderson, and the late Fred Fair, and of other series in the personal files of Charles F. Snow and of Eben G. Fine, both of Boulder, Colorado, disclose no evidence of slides of any magnitude since 1909. A few earlier photographs, because of limitations in resolving power and color sensitivity of the materials used, are nonrevelatory. Check observations, gleaned from the reports of mountaineering parties,

support the photographic evidence. Numerous avalanches are known to take place in this area during the winter, and rocks carried by them are eventually integrated into the rock glacier, either directly, or after assimilation by Fair Glacier. The importance and magnitude of avalanche action in this area, although believed great, are still uncertain because of its virtual inaccessibility in winter, when field conditions often approach those reported from the Antarctic (Ives, 1938b, 1940).

Evidence of motion.—Motion of Fair Valley rock glacier, although quite rapid, geologically speaking, is so slow, in terms of human chronology, that indirect, rather than direct, observations are necessary to show its existence. To an ordinary observer, this structure is as changeless as the surrounding cirque walls. Careful comparison of photographs taken at different times from the same station, however, shows plainly that some of the specific features of the rock glacier are differently shaped, or differently located, in the various pictures. Theoretically, the rate of motion of the structure could be determined by scaling the photographs. Actually, because of unavoidable photographic limitations, this has not been found possible. Despite these shortcomings, the photographs do show that the Fair Valley rock glacier is in motion, and that this motion is neither regular, nor uniform throughout its mass.

Traverses across the upper surface of the structure disclose that the equilibrium of most of the larger components is definitely unstable, particularly at the lower end, where the gradient changes sharply. The slight added load of a man walking across the rock glacier is sufficient, at times, to send several tons of debris crashing down the steep lower face into Triangle Lake.

Although the supply of water to this structure is quite constant from season to season, and has a rather uniform diurnal variation, resulting from the local daily storm cycle (Ives, 1938c), the positions of the various interstitial ponds, and their water levels, change from time to time, undoubtedly as a result of relatively minor motion of the rock glacier. The turbidity of the water in these ponds varies irregularly, each interior drainage being an independent variable.

Actual motion of this rock glacier has never been seen, so far as can be determined, but it has been heard on several occasions. During the early fall of 1932, while camping in Fair Valley, about half a mile below Triangle Lake, the writer was awakened by a series of crashings, rumblings, and earth tremors of such magnitude that the sudden collapse of several major peaks was suspected. Investigation, the next morning, disclosed no apparent changes in local topography. Later in the day, a visit to Triangle Lake, made in the course of other work, disclosed that

the water had changed from its usual glacier milk color to a dirty yellow, and was heavily loaded with mud. Further investigation showed that a number of the larger components of the rock glacier had rotated slightly, that some of the interstitial ponds had changed level and become very muddy, and that some of the coarse alpine grass had been uprooted. Forty-eight hours later, after two of the usual daily storms, no evidence of rotation could be found on the rocks, the interstitial ponds were again clear, and the uprooted clods had sagged back into position. Later, inquiry at the Regis College seismic laboratory, in Denver, showed that the vibrations set up by this shifting of the rock glacier left no observable seismograph trace.

On two other occasions, in 1933 and 1936, similar noises in the night were noted by the writer, and similar slight changes in the rock glacier were found. From a number of reports by mountaineering parties, rangers, and trail crews, it is concluded that three or four such minor readjustments take place during an average summer. Rather interestingly, all of them are reported to have occurred between 10 p. m. and 4 a. m.

HITCHENS' GULCH

General description.—Hitchens' Gulch rock glacier (SW½ sec. 35, T. 6 N., R. 76 W. Rocky Mountain National Park quadrangle), the largest and most representative rock glacier in the area studied, typifies the mature stage in the development of these structures. Located at the head of Hitchens' Gulch, a minor glaciated valley tributary to the Colorado River from the west, this structure is about 9500 feet long, nearly 2000 feet wide, and approximately 200 feet thick (estimated from extrapolation of valley profile). The upper surface of this structure is very flat, having a slope of about 8 degrees; the slope of the nose is about 30 degrees in most places. The normal angle of rest of talus in this valley is about 28 degrees. Relations of talus, rock glacier, and bedrock are shown diagrammatically in Figure 3. A view of the upper part of the structure, showing valley topography, constitutes Plate 2.

In the Hitchens' Gulch rock glacier the longitudinal ridges described by Capps (1910), Howe (1909), and others are fairly well developed, as is the steep outer boundary noted by Howe. In fact, Howe's photograph of the outer boundary of the rock stream in Pierson Basin (Silverton quadrangle, Howe, 1909, Pl. XI) could easily be mistaken for a view of the southern margin of the Hitchens' Gulch rock glacier, just below Lake of the Clouds. (An unnamed lake in NW¼, sec. 2, T. 5 N., R. 76 W., Rocky Mountain National Park quadrangle.)

The absence of a moraine of Neva age in this valley was noted. It appears that the valley was glaciated in Neva time, and that the rock

glacier overrode and assimilated the moraine of that age. Should the front of the rock glacier advance a few hundred feet, it would come into contact with the Monarch C moraine, of late Wisconsin age.

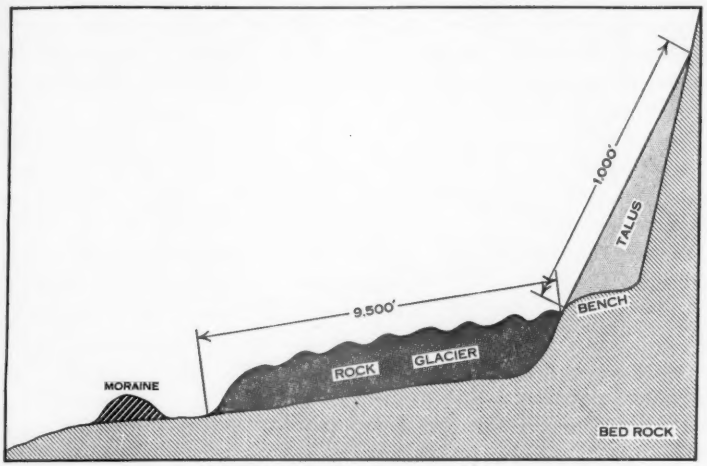

FIGURE 3.—*Longitudinal section of Hitchens' Gulch rock glacier*

Component fragments.—Components of the Hitchens' Gulch rock glacier are, in general, more uniformly sized than those of the Fair Valley structure, the average diameter of surface rocks being about 2 feet. A very few enormous fragments were noted, the largest measured being a nearly cubical mass of granite slightly less than 18 feet in diameter. Small amounts of rock flour settle out of the water running from the lower end of this structure. The absence of very fine material is somewhat surprising, for the crest of the range in this vicinity is veneered with volcanics, which weather rapidly into very small particles. As in the other rock glaciers nearby, the average size of the components decreases rapidly with depth.

Interstitial conditions.—The interstices of the Hitchens' Gulch rock glacier extend to a depth of more than 10 feet, as determined by actual measurement, and probably to depths exceeding 30 feet, as estimated by dropping pebbles and listening to the rebounds. In the lower part of the structure, the spaces between components are filled with gravel, sand, and organic material. Compacted rock flour is absent from accessible parts of the rock glacier, and, if present at all, fills only the lowermost interstices.

With the exception of a very small area near a minor permanent snowbank at the head of the gulch, no interstitial ice was found in this structure. It seems probable that very little ice is present, for the stream draining from the rock glacier is relatively warm, and lacks the characteristic taste of meltwaters. The small size of Hitchens' Creek, and the constancy of its flow, suggest either that its source is not in melting interstitial ice, or that the ice is at great depth, where it is not influenced by diurnal temperature changes.

No interstitial ponds were found in the accessible parts of this structure. Undoubtedly some are present in the center of the valley, under many feet of detritus, but their size and number are problematical, and need not be great.

In some of the larger cavities in this rock glacier there is a down draft, indicating the presence of open channels of some magnitude. According to theory, there should either be blowholes through which this indrawn air escapes, or a reversal of the currents at regular intervals. No evidence of reversal was found, and the number of blowholes located was small. However, whether the internal air circulation of this structure is a pulsation (diurnal or seasonal), or a continuous convectional draft, its existence supports the other evidence for open interstices.

Alimentation.—Most of the new rock supplied to the Hitchens' Gulch rock glacier comes from the overflow of a series of talus cones that line the valley walls near the head of the gulch. These talus cones rest on a rock bench, a remnant of an early Wisconsin valley floor, cut through, but not entirely removed, by later Wisconsin glaciation. Relations of talus cones to the rock glacier are shown in Figure 3, and in Plate 2. A continual noisy rain of rocks from the valley walls is a notable feature of upper Hitchens' Gulch. Unknown amounts of rock may also be added to the rock glacier by avalanches, which are probably numerous in this area, as evidenced by well-developed avalanche chutes on the upper walls of the valley. Dr. François E. Matthes of the United States Geological Survey explained these flutings on the upper walls of this and other valleys (personal communication, 1939).

Evidence of motion.—The Hitchens' Gulch rock glacier, like that in Fair Valley, moves so slowly that to a casual observer it appears changeless. Although the gulch has been known as an area of bad footing for more than half a century, the existence of the rock glacier as a discrete geologic structure apparently has not been recognized heretofore, and during this study no observations of any value were found in geological literature, mountaineer's reports, or photographic files.

FAIR VALLEY ROCK GLACIER

Glacier Extends Diagonally Downward from Fair Glacier, (Upper Right), to Triangle Lake (Lower Left). Note steep, but normal, talus just beyond Triangle Lake. Photographed in September 1932.

HITCHENS' GULCH ROCK GLACIER

Occupies Valley Floor and is Fed by Overflow from Talus Cones on far Wall of Valley, just below Vertical Ice-cut Cliff. Avalanche chutes are prominent just above the same cliff. Photographed in August 1939.

Like the Fair Valley structure, the surface rocks of the Hitchens' Gulch rock glacier are rather unstable, most of them being "rocking stones" which are quite difficult to walk across.

Absence of interstitial ponds, and of a terminal lake, inhibit detection of minor motions of the rock glacier. The steady production of rock flour, however, as evidenced by its continual presence in the stream draining the rock glacier, suggests that pulverization of rock may be taking place somewhere within the structure.

No actual motion of this rock glacier has been seen by the writer, nor has it produced any audible sounds during his visits. As both the writer's visits to the area, and shiftings of the rock mass, are irregular and independent variables, nonobservation of these sounds is not valid evidence that they do not occur.

Investigation of the rock mass near the lower end of the rock glacier discloses areas that are definitely static, there being large trees, perhaps as much as 200 years old, growing upright and unsplit among the rocks. Nearer the center of the valley, the interstitial fill is undermined in places, and cracked in others, indicating minor motion. As these cracks heal each year, as a result of frost action, followed by mudding of the fill when the snow melts, these appear to be the best evidence that the structure is not static. A number of stunted and misshapen trees, growing in the interstitial fill near these cracked areas, are also suggestive of motion, but their proximity to timberline, and their exposure to the violent winter winds in this area, make such evidence undependable.

HAGUE CREEK VALLEY

General description.—The Hague Creek Valley rock glacier (In the center of sec. 4, T. 6 N., R. 74 W., Rocky Mountain National Park quadrangle), a thoroughly unimpressive structure perched on the north wall of Hague Creek Valley, about 2 miles above "Cabin 13", at the junction of Hague Creek with the South Poudre River, is typical of the senile stage in the development of a rock glacier. This structure is approximately 2000 feet long (the lower end merges with a moraine, is tree-covered and somewhat hard to define), 500 feet wide, and more than 100 feet thick. The average slope of the upper surface is about 20 degrees, but this surface is broken into a series of steps, the treads being nearly flat, the risers sloping quite steeply. A longitudinal section of this rock glacier is shown in Figure 4.

It is notable that this structure occupies a portion of the valley that was glaciated in Monarch time, but was ice-free during the Neva glacial stage.

Component fragments.—Rocks composing the Hague Creek Valley rock glacier are smaller than those in the others described, few being

more than 4 feet in diameter, and most being not over 8 inches across. More fine material is present in this structure than in any of the others, particularly in the treads of its steplike surface. Many of the surface rocks have been split by frost action, and new surfaces are being created

FIGURE 4.—*Longitudinal section of Hague Creek Valley rock glacier*

on them by exfoliation, suggesting that the sizes cited do not represent original values.

Interstitial conditions.—The interstices are filled with gravel, sand, mud, and organic material in most places, open cavities being present only in the risers of the steplike upper surface, and these cavities are neither large nor numerous.

No interstitial ponds are visible, nor is the interstitial filling wet at the surface, but seepages from the lower margin of the structure suggest that it acts as a reservoir. The water drainage is warm, suggesting absence of interstitial ice.

Alimentation.—Formerly this rock glacier was fed by debris falling from a cliff above it. At present, the head of the structure has migrated, away from the cliff, and no more accretions of rock are probable from

this source. Between the head of the rock glacier and the cliff a normal talus cone has accumulated (Figure 4).

Evidence of motion.—All field evidence indicates that the Hague Creek Valley rock glacier is no longer in motion, and that it has not been in motion for several centuries at least. Large trees, estimated to be about 300 years old, growing upright in the interstitial fill in the treads of the surface, support this conclusion, as does the absence of rock flour from the seepages at the base of the structure.

This senescent rock glacier somewhat resembles an ancient, but slightly atypical, rock fall deposit. Comparison of details of this structure with those of known post-Monarch, pre-Neva rock falls, however, disclosed specific differences. Rock falls, in general, proceed not only to the bottom of the valley (or to a valley-wall bench), but overshoot, throwing numerous erratic blocks many feet up the opposite valley wall. The Hague Creek Valley structure not only did not reach the valley floor, or other point of decreased gradient, but also lacks the erratics characteristic of rock falls. Polished rock surfaces adjacent to known rock falls of this age are deeply marked by percussion spalls, the large star-shaped surface markings, closely resembling bullet scars, resulting from the impact of erratics. A type case of such a rock fall is located on the south wall of Landslide Basin (NE¼, sec. 3, T. 3 N., R. 74 W., Rocky Mountain National Park quadrangle). Well-developed percussion spalls are plainly visible on the north wall of the East Inlet (NE¼, sec. 18, T. 3 N., R. 74 W., Rocky Mountain National Park quadrangle) in the deep, ice-cut trough trending west from Lake Verna. These spall marks, in this area, are still visible after time lapses estimated from glacial chronology to be in excess of 5000 years. No such marks were detected near the Hague Creek Valley structure. Conditions on the north wall are such that a rock fall would normally proceed to the valley floor, throwing erratics many feet across the outwash material forming it. As the details of this structure all indicate former slow motion, measurable in very minute fractions of the velocity of free fall, it is classified as a senescent rock glacier, rather than as a rock fall deposit. Similarities here noted suggest that many ancient and static rock glaciers may have been misclassified as rock fall deposits.

ATYPICAL STRUCTURES

GENERAL STATEMENT

Throughout the area studied, and in all adjacent mountain ranges, are numerous detrital masses, which are neither typical rock glaciers (Pl. 1), nor normal talus cones. These appear to represent all gradations from

normal static talus to true rock glaciers. To them, the *omnium gatherum* classification, creeping talus, will be given, with the full realization that the term is indefinite.

Most of these structures resemble normal talus through much of their length, but have a slight mound at the bottom, or a creep ridge elsewhere, interrupting the normal regular surface profile. Ordinary evidence of motion is absent in these creeping talus cones, and the small amount of evidence fortuitously discovered in the Arikaree talus suggests that this motion is extremely slow, and only detectable, in most instances, by repeated detailed instrumental surveys, accurate to fractions of a foot.

Field studies indicate that a basal irregularity in a talus is not necessarily a product of creep, but may also be produced by the partial burial of a navigation ridge under later talus accumulation, as described by Behre (1933, p. 632). In the Southern Rocky Mountains, where there has been a relatively recent minor glaciation (Neva stage), it seems probable that most of the basal irregularities in talus cones are not creep ridges, but reflections of buried or partially buried nivation ridges, and should be classified as such in the absence of conclusive evidence of motion. Satisfactory criteria for prompt determination of the origin of basal irregularities in talus seem to be lacking.

THE ARIKAREE TALUS

General description.—The Arikaree talus (NE¼ SE¼ sec. 14 and NW¼ SW¼ sec. 13, T. 1 N., R. 74 W., Rocky Mountain National Park quadrangle [2]) descends in typical conical form from the col between Arikaree and Albion Peaks to the floor of the Green Lakes Valley, 750 feet below. The base of the cone is approximately 400 feet across; its surface slope averages 29 degrees. This slope is one of the routes to the "peneplain" from which ascents of Arikaree Peak are made, and its motion was determined from continual shiftings of strategic footholds on the trail across the detritus.

Component fragments.—Rocks composing the Arikaree talus are of all sizes from about 2 inches to 5 feet in length. Many of those on the surface are fairly large slabs, with the longest dimension about 2½ feet and the shortest about one foot. Most of this material is granite and granite-gneiss, but a few schist boulders are present.

Interstitial conditions.—Accessible interstices in this talus are open, but the cavities are not large. Running water, from the snowbanks above, can be heard at no great distance below the surface, and issues in a series

[2] The names Kiowa Peak and Albion Peak have been transposed in this map.

of small trickles at the base. During most of the short mountain summer this water is clear, but for a few days in early spring (mid-June at this altitude), some fine silt is carried.

Alimentation.—The Arikaree talus is fed from two sources. Fragments of rock rain onto it from the glacially oversteepened north flanks of Arikaree and Albion Peaks. Accretions from these sources are numerous, but small, and many of the rock fragments are so deeply weathered that they disintegrate on landing. Slabs of rock, of various dimensions, slowly creep onto the talus from the gently-sloping "peneplain" south and east of Arikaree Peak. This type of creep, the result of the combined actions of frost and gravity, is adequately described in existing literature, which is reviewed by Sharpe (1938, p. 27-28, Fig. 3 *A*, p. 110). Available evidence suggests that talus accretion from this source is measurable in terms of pounds per year.

Evidence of motion.—Even a detailed inspection of the surface of the Arikaree talus conveys an immediate impression of great stability. Surface rocks, in general, are so firm that they do not teeter under the weight of a man carrying a heavy pack. No change in profile or appearance has been noted in this structure during the 8 years it has been under observation; none are reported to have taken place since before the beginning of the century. The usual clearness of the water flowing from the base of the structure indicates absence of internal milling. The presence of rock flour during the early spring, when runoff is great, suggests a general flushing of the interstices, rather than motion of the structure. Were it not for the rotation of some of the surface rocks, this structure could be classified as an extremely stable talus.

Rotation of one component, which formerly was an important foothold in the trail, is shown diagrammatically in Figure 5. The dimensions in this figure are estimated, the probable error being plus or minus 2 degrees and plus or minus one inch. It will be noted that this observed rotation is backward, but, as shown by Sharpe (1938, p. 67-74), such backward rotation is quite common. Theoretical consideration of the vertical shrink of a talus undergoing compaction suggests that a tendency to backward rotation of components is normal, and that any surface block rotating forward is not partaking of the motion of the structure as a whole, but has escaped and is merely rolling down the sloping surface.

Observed rotation of surface blocks in a talus cone is far greater than that called for by theoretical considerations, but only a few of the blocks rotate. This seeming discrepancy can be explained as the result of inter-

action of components. Relief over a discrete area is obtained by considerable rotation of a single component, rather than by a slight rotation of each. This concentrated relief is similar to the rupture of a floor after considerable wetting—every board is swelled, but only at rather widely spaced intervals is an adjoining pair arched upward.

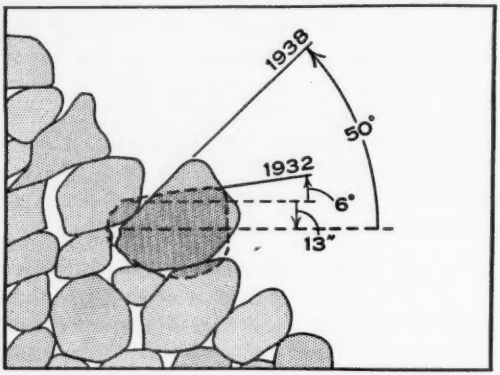

FIGURE 5.—*Rock rotation in the Arikaree talus*

STRUCTURES RESEMBLING ROCK GLACIERS

A number of high cirques in the Southern Rocky Mountains contain masses of detrital material bounded by morainelike ridges. These structures closely resemble rock glaciers, and may be mistaken for them even after rather careful study. As there is some rearrangement of components in any detritus subject to periodic freeze-and-thaw, there will be some evidence of motion (usually slight) in these structures.

These pseudo-rock glaciers appear to be formed when talus banks up behind an abandoned moraine in a cirque. In general, if the boundary ridge of a suspected structure is located in the logical position for a moraine which is otherwise unaccounted for, the classification of the structure, however much it resembles a true rock glacier, should be regarded as doubtful. Relatively old moraines can be distinguished from the marginal ridges of active rock glaciers by differences in weathering, but moraines of the "little ice age" (Neva glacial stage) are so fresh that this criterion is of little value, and only successive observations, of sufficient accuracy to detect motions of the order of a few inches a year (or less), will show whether the structure is a banked morainal loop or a rock glacier.

ROCK GLACIER MOTION

GENERAL STATEMENT

Theories of rock glacier motion have been advanced by Howe (1909), Capps (1910), Tyrrell (1910) and Hole (1912), and the subject of rock glacier motion has been rather exhaustively reviewed by Sharpe (1938, p. 42-46).

Howe (1905) ascribed the structures to detrital deposition upon snowbanks and receding glaciers, a process somewhat analogous to the formation of nivation ridges (Behre, 1933), but later (1909) reached the conclusion that they were enormous rock falls, the surface wrinkles being attributed to impact effects. Although this view is not given wide credence today, certain structures such as the rock glaciers on La Veta Peak, Colorado (Patton, 1910; Sharpe, 1938, p. 44-45) may owe their origin, in part at least, to rock falls. Similar structures, on a small scale, may be produced experimentally by dropping shovelfuls of sand onto a sloping surface from a height of several feet. As was first noted by Heim (cited by Capps, 1910, p. 370) in 1882, flow structures are remarkably similar, whether produced in a few minutes or a few centuries, and from this similarity has come considerable confusion between the various structures produced by flow. It is notable that the Slumgullion Mudflow (*see* Slumgullion Gulch, San Cristobal, Colorado, quadrangle), when seen from a distance, has the same general appearance as a rock glacier (Howe, 1909, p. 50, Pl. XX, B.).

Capps attributes rock glacier motion to periodic freezing and thawing of interstitial ice, a view supported by Hole, and in general agreement with the greater part of the evidence assembled by Sharpe.

Tyrrell believes that rock glacier motion is caused by periodic freezing of spring water within the detritus, but illustrates his paper with a photograph (Tyrrell, 1910, Fig. 1) of a structure that would today be classified as a rather active creeping talus cone.

Although rock glaciers and related structures are found in many areas, in which environments and conditions are somewhat dissimilar, certain features are common to all of them. For a rock glacier to exist in a given locale, there must be a plentiful supply of detritus, and a set of conditions such that this detritus does not form normal talus. As both rock glaciers and normal talus are found adjoining in the same areas, it appears that the factors producing rock glacier motion are not intrinsic to the components, but are the result of forces acting through the interstices and along the planes of contact.

CLIMATE

Climatic conditions in the region above timberline have been discussed by Capps (1910), Daly (1905), Matthes (1900), and many others. De-

tails of the climate of the Colorado Front Range have been reported at some length by Ives (1938b, 1938c, 1940), and the effects of alpine climate pertinent to studies of talus accumulation and motion have been reviewed by Behre (1933) and Sharpe (1938, p. 33-46).

From timberline to the summit of the Front Range, climatic conditions definitely favor both rapid production of detritus and instability of talus accumulations. These conditions closely approach those reported from similar mountain regions elsewhere.

Surface temperatures, during many days of the four months annually that the region is free of snow, vary from slightly below freezing (about 2 a. m.) to somewhat over 90 degrees F. (about 1 p. m.). This variation, creating a periodic freeze-and-thaw in the outer parts of exposed rocks, causes rapid disintegration, which leads to extensive talus formation. The daily storm cycle (Ives, 1938c), which thoroughly wets exposed rock surfaces just before the evening fall in temperature, accelerates disintegration. During the winter, when air temperatures fall far below zero, exposed rock surfaces are warmed to temperatures above freezing by the intense sunlight of the high altitudes, so that disintegration, while probably not as rapid in winter, still continues.

Within a moist or wet talus, periodic freezing and thawing supply thrusts which initiate or assist creep. In this area it appears that interstitial ponds exposed to outside air freeze and thaw many times during the year, but those some feet below the surface, having less exposure, freeze and thaw annually. This leads to differential motion of the various zones of the structure, a factor of some importance in the production of surface irregularities.

In the Fair Valley rock glacier, the diurnal freeze-and-thaw is of considerable importance, there being numerous interstitial ponds exposed to minor local temperature changes. Should thrusts be uniform throughout the mass, fewer surface irregularities would be present, and such is the case in the upper part of the Fair Valley structure, where interstices are filled with rock flour, which tends to absorb such thrusts as are produced by freezing of its moist surface, and to insulate subsurface mud from the effects of short-lived temperature variations.

In contrast, the Hitchens' Gulch rock glacier, which is substantially dry in its upper parts during the summer, can be subject only to seasonal freeze-and-thaw, which should lead to slower motion, a conclusion in accord with field observations.

ROCK TEXTURE

Although the angle of rest of a talus cone seems to be largely a function of component size (Behre, 1933, p. 623-626), certain properties of

the components, here loosely termed "texture," may be of some importance in producing the instability necessary for both talus creep and rock glacier motion.

The mechanical strength and coefficient of friction of many rocks decrease as they absorb or adsorb water. Accordingly, a talus cone may be relatively stable during dry seasons, and somewhat unstable when wet. Calculations from rather unsatisfactory data suggest that the maximum changes in rock properties possible from wetting are just adequate to produce creep in a talus cone that is on the ragged edge of stability when dry.

The properties of the products of weathering and mechanical disintegration of components of a talus cone may also have important effects on its ultimate stability. Should the products of disintegration be angular, so that their presence between components would increase friction, the structure would tend to become more stable with age; but if these disintegration products tended to decrease friction between components, the structure would become less stable as it became older.

Porosity of components, in detrital structures having adequate water supply and subject to periodic freeze-and-thaw, may be of minor importance, for the life of a porous rock, under such conditions, all other factors being the same, is shorter than that of a relatively impervious one. The collapse of a buried component of a detrital structure leads to internal readjustments, with resultant internal milling, shearing of corners and thin edges of other components, and other complex interactions which proceed outward from the source of local instability until damped out by friction or stopped by the bridging of an opened cavity.

SLIP PLANES

The possible existence of slip planes beneath rock glaciers and related structures has apparently been given little consideration by most workers. Capps (1910, p. 364) considers "the shape and grade of the floor over which it moves" an important factor governing the rate of movement of a rock glacier. Salomon (1929) discusses the effect of glacial clays and muds beneath a rock glacier, and concludes that saturated rock flour may act in much the same manner as a "partially fluent zone" (Sharpe, 1938, p. 71).

Every valley in the Colorado Front Range that contains a rock glacier or related structure was glaciated several times during the Wisconsin. Hence, it may be assumed with some confidence that there is a smoothed rock floor, capable of acting as a slip plane, beneath the upper part of each rock glacier. The lower parts of the more extensive structures, such

as that in Hitchens' Gulch, are overriding ground moraines and outwash deposits, which may be acting as partially fluent zones.

Theory suggests that a partially fluent zone under heavy load offers more resistance to detrital motion than does a polished rock surface, all other factors remaining the same. Under these conditions, after a rock glacier has traversed a certain extent of ground moraine and outwash, the force necessary to overcome basal friction exceeds that necessary to produce shear and overriding within the structure, and unless the rock glacier stalls because of friction, it will acquire transverse ridges. Such ridges are common in the lower parts of many glaciers, both in this region and elsewhere, but are absent from the upper parts of the structures, where longitudinal ridges are present. Although conclusive evidence is lacking, it seems reasonable to assume that there is an approximate coincidence of the point where the surface ridges change from substantially longitudinal to substantially transverse with that where the slip plane gives way to ground moraine and outwash.

LUBRICATION

Possible lubricants of detrital structures at high altitudes are water, ice, mud, and algae. Although dry sand and dust may, under certain conditions, act as lubricants, they cannot remain dry, in the environment under consideration, for long enough to be of any importance.

In structures composed of large fragments, where capillary effects in the interstices are negligible, water alone is an ineffective lubricant. Only where water has some definite effect on the components, as in reducing the coefficient of friction by being adsorbed, does it become an important lubricating agent.

The action of ice in a detrital structure may be likened to the behavior of a cheap, heavy oil in a motor. Under moderate pressures, at low velocities, it is a fairly good lubricant. Where velocities would otherwise be high, ice, because of its viscosity, acts as a damping agent. Where pressures are high, ice squeezes out of the zone of contact, redeposits where the pressure is less, and in no way assists motion.

The effects of mud on a detrital structure are varied, and usually complex. When sufficiently liquid, muds buoy up the components, reducing the forces holding them together, and hence lessen the effective friction. When drier, most muds inhibit motion. As pressure increases, muds tend to pack, reducing whatever lubricating qualities they may have.

Salomon (1929) discusses the lubricating qualities of glacial clays, as applied to rock glacier motion. Under certain conditions, glacial clays are good lubricants, but in many instances these so-called glacial clays are actually pulverized rock, a rather good abrasive, and no more useful

in the lubrication of a rock glacier than is emery powder in a bearing. In the Colorado Front Range area certain glacial deposits are true clays, and certain other deposits of similar-appearing materials are settled-out rock flour. The loose application of the term clay to all of these deposits, with no regard for their kaolin content, is unfortunate and misleading.

Algae, diatoms, and other small aquatic flora are common in streams flowing from detrital structures, and may be found in considerable numbers in some interstitial ponds. Although no detailed studies of lubricating qualities have been made, an algae-covered surface that is slick to the touch is still slick under pressures approximating 10,000 pounds to the square inch (steel to granite, with algal material between).

SUMMARY OF MOTIVATING FORCES

In the foregoing discussion, which is by no means exhaustive, a few of the factors in rock glacier motion have been briefly touched upon. Of these, the climatic factors appear to be of major importance in producing not only detrital instability but rapid detrital accretion. Secondarily, slip planes beneath the structures seem to be of importance, although their presence is inferred from glacial mechanics, rather than demonstrated. The effects of rock texture appear to be negligible in most instances, but the effects of thermal disruption of saturated porous components may, in some structures, be the cause of creep. The actions of supposed lubricants are somewhat problematical under conditions prevalent within rock glaciers, particularly as water is an ingredient of all of them. Algae appear to be a hitherto unconsidered lubricating agent which needs further study before its effects can be evaluated.

ROCK GLACIERS IN ADJACENT AREAS

Reconnaissances in all of the eighteen major mountain ranges in Colorado (Ives, 1938c, map, p. 693) disclose some structures in each range that are not normal talus cones, and many structures that are probably active rock glaciers. These reconnaissances, augmented by rather extensive studies of the photographic files of the United States Geological Survey, the National Park Service, the Colorado Mountain Club, the San Juan Mountaineers, the University of Colorado Museum, and the personal collections of William H. Jackson, Charles F. Snow and H. L. Standley, lead the writer to conclude that in these ranges about one talus cone in fifteen shows definite signs of instability. Of the structures that are apparently unstable, perhaps one in twenty is a rock glacier.

The San Juan Mountains, of southwestern Colorado, have become a type area for rock glaciers originating in cirques, largely as a result of the

work of Howe (1909) and Hole (1912). Numerous rock glaciers here, and in the adjacent San Miguel Range, remain unstudied.

Many small structures resembling diminutive rock glaciers are present in the Sangre de Cristo, Sierra Blanca, and Culebra ranges of south-central Colorado. Just east of the Sierra Blanca is Veta Peak (SE¼ of Huerfano Park quadrangle, near Muleshoe Station, Denver and Rio Grande Western Railroad), a type area for rock glaciers apparently unrelated to glacial phenomena (Patton, 1910).

Many well-developed rock glaciers are present in the West Elk Mountains of central Colorado, particularly in the vicinity of Snowmass Mountain. General properties of the rock of this area and climatic conditions suggest that these structures are probably in rapid motion, and that this region should be studied in detail, so that definite knowledge of the rate of motion of the structures can be secured. Superlative photographs of parts of this area, with definition adequate for use in such studies, were taken in 1872 by William H. Jackson, and are now in the photographic files of the United States Geological Survey. Rephotographing of these areas, from the stations used by Jackson 68 years ago, and comparison of the old and new photographs, should show definitely whether these structures are in motion, and give a close approximation of the rate of motion.

Somewhat atypical talus accumulations are present in the Rabbit Ears, Kenosha and Tarryall, Wet, Gore, Flat Tops, White River, and Collegiate (Sawatch) mountain ranges, but no rock glaciers in these ranges are known to the writer. This should not be interpreted as a statement that no such structures are present, for there has been little detailed study of these regions.

The Medicine Bow and Never-Summer Mountains, which adjoin the Front Range, contain many small structures that may be either diminutive rock glaciers or rapidly creeping talus cones. The slopes of Pike's Peak support a number of slightly atypical talus accumulations, but no well-developed rock glaciers.

In the Mosquito Range, in which detailed studies have been made by Capps (1909) and Behre (1933), several rock glaciers are present. Two of these, because of their accessibility, probably are worthy of detailed investigation. One, in Buckskin Basin, at the head of Buckskin Gulch Leadville quadrangle, (lat. 39° 20′, long. 106° 7½′) appears to be greatly influenced by avalanches, which periodically disrupt mining operations in the area. The other, on the north flank of Mt. Lincoln, appears to be a typical rock glacier occupying a cirque, and is clearly visible from the summit of Hoosier Pass. Drainage from this structure (located in Leadville quadrangle, lat. 39° 21¼′, long. 106° 5′) is tributary to Platte Gulch.

Several "talus glaciers" on the western slope of this range are described by Emmons (1927).

SUMMARY AND CONCLUSIONS

Field investigations here reported indicate that rock glaciers and related structures are by no means uncommon in areas recently eroded by alpine glaciers.

Motion of the structures studied in detail appears to be caused by the thrusting of interstitial ice, there probably being two cycles of thrusting—a diurnal cycle occurring in the surface layers during the summer, and an annual cycle affecting the whole structure.

Surface irregularities appear to be caused by nonuniform motion of the components, produced by changes in basal friction, or by differential motions of various layers.

The presence of slip planes or partially fluent zones is inferred from the locations of the structures in recently glaciated valleys.

Lubricants seem to be of relatively minor importance in the structures studied, but further investigation of suspected lubricants, particularly algae, seems desirable.

Throughout this investigation, the need for more data, most of it obtainable only by detailed and long-continued field investigations, has been apparent. Although the findings here reported substantially confirm the explanation of rock glacier motion given by Capps (1910), confirmed by Hole (1912), and reviewed by Sharpe (1938), certain minor features, such as interstitial ponds, suggest that Chrystocrene action, described by Tyrrell (1910), may be a minor motivating force in some structures. The effects of rock falls and avalanches, suggested as a major motivation by Howe (1909) and as a minor one by Patton (1910), should not be disregarded prior to careful evaluation, particularly as avalanche deposits melt away from many of these structures before they are accessible to a man afoot.

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ORIGIN OF THE MUCK-SILT DEPOSITS AT FAIRBANKS, ALASKA

BY RALPH TUCK

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ABSTRACT

The extensive gold-bearing gravel deposits in the Fairbanks district are covered with a muck-silt overburden varying from a few feet to over 200 feet thick. This overburden is economically important as (1) Its thickness is often a limiting factor in determining whether the underlying gold-bearing gravel can be profitably mined, since with large-scale mining methods it must be entirely removed, and (2) The difficulty and cost of exploration have been increased, as the overburden has concealed many of the gold-bearing channels.

The overburden is of two types—muck and silt. The muck, which is frozen, occupies the present stream valleys and lies unconformably upon, or grades into, the silt, which is unfrozen and occurs on the lower hill-slopes. The inorganic material

in the two deposits is essentially the same, but the muck contains about 50 per cent ice and vegetable material as well as an abundant Pleistocene fauna.

The origin of the muck and silt has long been obscure because of poor exposures and the complications of sub-Arctic solifluction processes. It now seems certain that the silt is of aeolian origin, while the inorganic material of the muck consists principally of wind-blown particles that have been reworked by creep, sheetwash, and stream action. A large part of the interior of Alaska is unglaciated; the deep muck-silt deposits occur in this unglaciated area and are probably a result of Pleistocene glaciation in nearby localities, as conditions were then ideal for aeolian deposition.

INTRODUCTION

The Fairbanks district is known principally for its placer gold production, which since its discovery in 1903 has had a total value of over $110,000,000. In 1937 the value of the placer gold mined was about $5,000,000, and the district has produced on a similar scale since then and probably will for many years to come.

The muck-silt deposits in the district, as well as in adjoining regions, have considerable economic significance, since they blanket a large part of the area containing the gold-bearing gravel deposits and, therefore, affect placer prospecting and mining operations.

Because of this overburden the cost and difficulty of exploration are greatly increased, as usually a great thickness must be penetrated before the stream gravel is reached, and, still more important, it has topographically obscured or concealed many of the old gold-bearing channels. Practically all the early placer gold exploration was carried on by sinking shafts through this muck-silt overburden into the underlying gravel, and hundreds of shafts were necessary to follow the pay channels. Many such prospecting attempts were abandoned either because of depth of overburden or the encountering of a flow of water upon reaching gravel. Present prospecting is done principally by churn drilling, so that in areas of deep overburden discoveries of gold-bearing gravel are occasionally made.

In the early days of the district, mining in the areas of deep overburden was done by drifting from the prospecting shafts. Under present conditions, in all large-scale and therefore open-cut operations, the muck-silt must be removed in order to mine the underlying gravel (Pl. 1); consequently, the thickness of overburden partly determines the mining cost and frequently is the limiting factor in determining whether the gravel can be profitably mined.

The origin of this muck-silt overburden, which has long been a controversial question, has been obscured by the mantle of vegetation and by the occurrence of weathering and soil-creep conditions peculiar to sub-Arctic climates. Only in recent years, through the stripping of large areas preparatory to dredging, has a more complete picture been revealed.

These deposits were first examined by the writer while employed as

geologist for the government-owned Alaska Railroad. Later, while employed by the United States Smelting, Refining, and Mining Company, many open-cuts were examined in various stages of stripping, and opportunity was afforded to study the prospecting records, principally drill

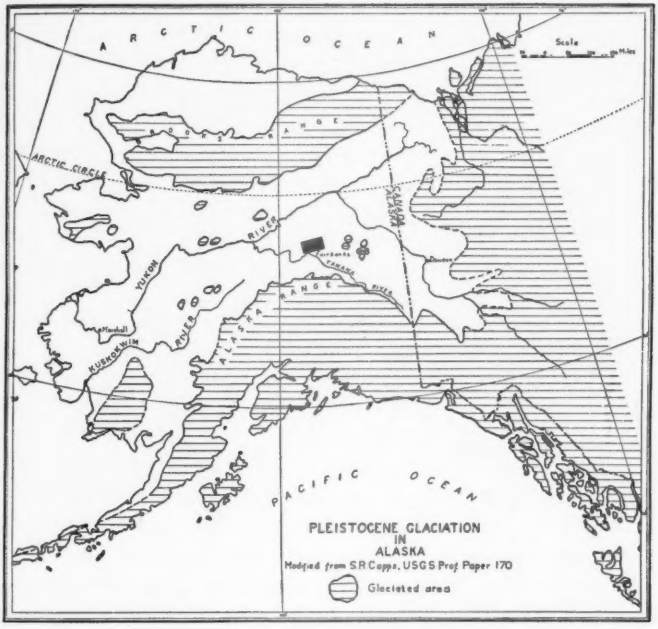

FIGURE 1.—*Map of Alaska showing location of Fairbanks district and glaciated and unglaciated areas*

logs from many holes put down over a period of 15 years. The members of the Alaskan Branch of the United States Geological Survey and the officials of the United States Smelting, Refining, and Mining Company have contributed many ideas and suggestions as to the problem of the origin of the muck-silt deposits.

GENERAL FEATURES OF THE DISTRICT

The Fairbanks district lies in the interior of Alaska (Fig. 1). Its extent is roughly determined by the creeks that carry placer gold in the vicinity of the town of Fairbanks, and it is about 40 miles long by 20 miles wide.

The Tanana River, one of the major tributaries of the Yukon River, borders the district on the south side. The placer creeks, all of which directly or indirectly are tributaries of the Tanana, drain an area of low rolling relief, a late mature topography, ranging in elevation from 450 to 2600 feet. Divides and ridges are roughly concordant in elevation, except for two prominences known as Ester and Pedro domes, which rise a few hundred feet above the general ridge level. South of the district are the gravel flats of the Tanana Valley, which in the vicinity of the town of Fairbanks have an elevation of about 450 feet and rise gradually to the flanks of the Alaska Range about 60 miles to the south. These gravel flats are not caused by the aggradation of the Tanana River alone but are in part outwash plains of many glacier-fed streams flowing north from the Alaska Range into the Tanana. Like the greater part of the Yukon Basin the Fairbanks district is unglaciated (Fig. 1), but during the Pleistocene glaciers probably extended out into the south side of the Tanana Valley to within 40 miles of it.

Quartz-mica schist is the dominant bedrock, but there are a few areas of intrusive granodiorite. The courses of the major streams are controlled by the regional northeast—southwest structure of the schists, but practically all the gold produced has come from creeks that flow radially from three distinct granodioritic areas.

Extensive gravel deposits occupy all the stream valleys and range in thickness from a few feet on the headward portions of the creeks to over several hundred feet on the lower portions close to the Tanana Valley. The Tanana Valley, which is the local base level of all the placer-gold streams, is filled to an unknown depth with alluvium. A few holes on the edge of the valley have been drilled to a depth of over 400 feet—50 feet above sea level—without striking bedrock. Reconstruction of bedrock grades of tributary streams also indicates that bedrock in the Tanana Valley at Fairbanks is at, or below, sea level, although 1000 miles from the ocean. The great fluctuations in the base level of the Tanana River have made the erosional history of the tributary streams extremely complex with a number of drainage changes and stream captures, but such changes are now obscured by alluvium, and muck and silt, and can only be determined by a great amount of deep drilling.

A large part of the creek gravel is frozen to bedrock, and this depth of permanent frost is sometimes as great as 200 feet. Throughout the interior of Alaska, a large area of permanently frozen gravel and subsoil coincides almost always with the unglaciated area, thereby suggesting a possible cause for this condition.

AERIAL PHOTOGRAPH SHOWING MUCK-STRIPPED VALLEYS ABOVE DREDGED AREA
(Courtesy of United States Geological Survey.)

FIGURE 1. ICE SILLS IN MUCK
(Courtesy of United States Geological Survey.)

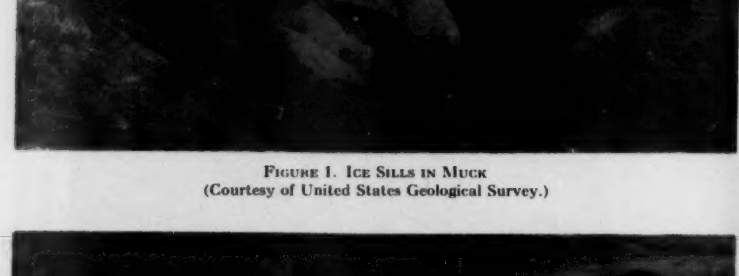

FIGURE 2. ICE DIKES IN MUCK
(Courtesy of United States Geological Survey.)

TYPICAL OCCURRENCES OF MUCK ICE

MUCK-SILT DEPOSITS

GENERAL STATEMENT

The term "muck" was early applied by the miners to an almost black, frozen, fine-grained sediment, containing a great amount of vegetable material, some vertebrate remains, and many ice lenses, and occurring in the present stream valleys and overlying the gold-bearing gravel. Locally, the term has been misused to include a silt that is light-colored, unfrozen, fine-grained, contains little or no organic material, and is found on the lower hillsides and also occasionally overlies auriferous gravel. Some confusion in the use of the terms "muck" and "silt" is inevitable, as gradations between the two are common. Practically all the creek gravel, as well as a part of the bedrock, is overlain by muck and silt. The deposition of the muck and silt has been the most recent geologic event in the district; all the streams, with the exception of the Tanana River, are flowing in it and in only a few places have they been able to cut through into the underlying gravel or bedrock.

MUCK

The muck covers the valley bottoms of all the streams in the district, with the exception of the Tanana River where gravel flats are exposed. From the surface, where it is covered only by the present-growing vegetation, it extends to depths ranging from a few feet on the headward portions of the valleys to 150 feet on the lower portions of the large streams. The base of the muck horizon that mantles the surface does not extend more than a few feet below the elevation of the Tanana Valley. However, deep drilling indicates that close to the Tanana flats there is a lower muck horizon separated by gravel from this upper surface one.

Uniformly fine mineral grains, of which more than 98 per cent will pass through 200-mesh screen but of which there are practically no clay sizes, make up the inorganic material in the muck. The mineral grains are subangular and are predominantly quartz and mica, with lesser amounts of feldspar, hornblende, rutile, garnet, epidote, and other accessory minerals.

The amount of organic material varies from a trace to almost 100 per cent—where it forms peat—but the typical muck contains from 20 to 50 per cent. Practically all the organic material has come from grasses, mosses, alder, spruce, willow, cottonwood, and birch—a vegetation identical with that growing at the present time. The rude bedding, imparted to the muck by the vegetation, is usually horizontal or at a low angle conforming with the slope of the valley, except where it is locally highly contorted by ice dikes. Much of the vegetable material

is in place, with the stumps of trees still embedded and upright; in one 20-foot section, six horizons of residual vegetation were plainly evident. In other localities, the heterogeneous accumulation of tree trunks and branches indicates that it was washed in. Occasionally there is evidence of soil flows.

FIGURE 2.—*Generalized section of muck on Engineer Creek, Fairbanks*

Ash layers several inches thick mantle some of the horizons of vegetation. Where the vegetable material of these horizons is in place, the inorganic material of the muck is always slightly oxidized, indicating that the surface was exposed to weathering for some time (Fig. 2).

Vertebrate remains—usually bones, but sometimes almost complete skeletons, and occasionally bones or skeletons with skin and flesh adhering—occur throughout the muck. These remains include living species such as moose, caribou, and many smaller types, and extinct species such as mammoth, mastodon, sabre-toothed tiger, super-bison, and camel. The occurrence of most of the remains indicates that they have been washed, or slid, into their present location, but a few are found in place. Usually where one fossil is found there are many, suggesting the possibility of a water-hole where they congregated or a boggy spot where they may have been entrapped. The finding of complete skeletons indicates that the remains are where the animals died, and, when flesh and hide are still adhering, that freezing and covering must have occurred shortly after death. Rodents' nests are common and are always found a few feet below the horizons of residual vegetation. Organic material is so

abundant in the muck that the odor from its decay is noticeable half a mile or more from the open-cuts.

All the muck is frozen, and three types of ice are recognizable: (1) One type is that inherent in the muck, occupying the pore spaces, which was formed contemporaneously with the muck itself. This type forms about 50 per cent of the total mass, which corresponds to the porosity of the inorganic material under close packing. (2) The second type of ice is in sills, a few inches to 10 feet thick, which conform with the bedding of the muck (Pl. 2, fig. 1). The sill ice contains a small amount of organic and inorganic material whose rude bedding conforms with that of the main muck mass. The sill ice is believed to be contemporaneous with the muck and to be caused by the freezing of the overflow of creeks in winter and the subsequent covering with silt and vegetation in the spring. A similar process may be observed at the present time. (3) The third type occurs as dikes of clear ice cutting across the sill ice and the bedding of the muck and distorting the latter (Pl. 2, fig. 2). The most probable explanation of the dikes is that they represent tension fractures in the frozen muck that have become filled with water—either ground or surface—which subsequently froze, thereby distorting the bedding of the muck.

The muck is always very dark because of the presence of frozen organic material, but upon exposure the vegetable material oxidizes, and the muck assumes a light color identical with that of the silt.

When the muck directly overlies gravel, the latter is always frozen. The contact between the two is usually sharp but sometimes is a gradational zone of interbedded muck, sand, and fine gravel. At the valley heads, the muck may lie directly upon bedrock, which is then usually frozen for several feet. The muck sometimes lies directly upon silt, in which case the contact is usually sharp and the silt is frozen back for 10 or 15 feet.

In the large valleys the muck has many irregular topographic features, such as small ridges and hummocks, and elliptical and circular depressions occupied by small lakes. The irregular depressions are probably due to subsurface thawing of the frozen material by circulating ground water and consequent subsidence of the surface. The elliptical and circular depressions are always surrounded by a muck ridge from 5 to 20 feet high, which is broken at one end. The perpetually frozen subsoil serves as an impervious layer, so that much of the ground water is under a considerable head. These depressions have probably been caused by hydrostatic pressure doming up the frozen layer, which subsequently collapsed upon the release of pressure, leaving a ridge around the depression that was breached at one end by the outflowing water.

SILT

The silt mantles the lower slopes of the larger hills and ridges, caps low hills and ridges, and ranges from a few to 300 feet thick. It extends down into the valleys but always gives way there to the muck. No definite upper limit of the silt exists; usually there is not much above an elevation of 1200 feet, but it has been found as high as 1800 feet. The lower limit is unknown; in one locality deep drilling has shown that it extends continuously from the surface to a depth of over 300 feet where it would be 150 feet below the elevation of the Tanana River.

The heavy cover of moss and vegetation obscures its areal extent, but the upper limit often shows topographically by a change in slope, as the bedrock slope is usually steeper than that of the silt. The extent and thickness of the silt diminish rapidly away from the Tanana Valley. Close to the Tanana it forms many long spurs, extending from the bedrock ridges and topographically concealing old drainage systems.

Typically it is light tan, unfrozen, and contains little or no organic material, although there are occasional gradations into the muck. Mineralogically and texturally the silt is similar to the inorganic material of the muck. Few silicified bones have been found in it. A faint horizontal bedding is sometimes present, but characteristically it has none and upon weathering it breaks into vertical faces. In some places the silt is slightly cemented with iron oxide but in general it is unconsolidated. In a few exposures small calcareous concretions occur at definite horizons. Minute tubelike casts are universally present in all the typical silt exposures.

Surface stripping preparatory to dredging operations at the confluence of Ester and Cripple creeks has afforded a 120-foot vertical exposure of silt, which shows three ash layers within a 25-foot zone. The ash zone is about 15 feet above the base of the silt, which overlies gold-bearing gravel. The volcanic ash layers, 2 to 6 inches thick, contain thin films of carbonaceous material. Except for local irregularities caused by compaction, they are horizontal (Fig. 3).

Where gravel underlies the silt, the contact is usually a 2- to 4-foot zone of interbedded silt, sand, and fine gravel, and, like the silt, the gravel is always unfrozen. On the hillsides the silt lies directly on weathered bedrock, which likewise is unfrozen, and the contact is sharp.

The silt, found principally on the lower hillsides, forms a good agricultural soil and supports practically all the productive farming of the district. Not only does the silt form a good soil but, because of its advantageous topographic position, it is well drained, receives a maximum amount of sunlight on south-facing slopes, and has a better possibility of escaping late frosts in the spring and early frosts in the fall.

RELATION OF MUCK TO SILT

A close relation of the muck to the silt is evident from the similarity of the inorganic material, the gradations in organic content, and the field associations, so that the origins are in part interrelated.

The similarity of the mineral constituents is very obvious, particularly as to size, as screen tests from the two deposits are practically identical,

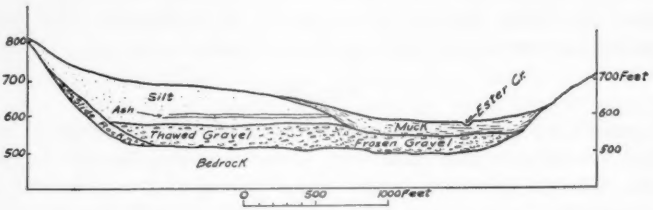

FIGURE 3.—*Section showing relations of silt, muck, and frozen and unfrozen gravel on Ester Creek*

95 to 98 per cent of the material passing through a 200-mesh screen, and the clay content being less than 3 per cent. No detailed petrographic study of the minerals has been made, but a cursory examination indicates that the common constituents are essentially the same. Megascopically this similarity is evident, for after weathering the muck cannot be distinguished from the silt.

Complete gradations in the amount of vegetable material from silt to muck have not been found in any single exposure. However, partial gradations can be seen, and in a number of exposures over the entire district almost any gradation can be found, ranging from silt without any vegetable material to the typical muck with about 50 per cent, and finally to peat with practically no inorganic material.

Actual field relations have in the past been difficult to obtain, as most of the mining operations are in the present creek valleys, so that practically all the open cuts are in muck, whereas the silt is principally on the hillsides. With the approaching exhaustion of some of the present creek deposits, more widespread exploration was undertaken, and auriferous channels topographically concealed by silt and muck were found. In the last 3 years, stripping of the overburden preparatory to mining operations on one of these concealed channels at the confluence of Ester and Cripple creeks has uncovered a number of excellent exposures of the two deposits. In all these the silt is older than the muck, as the latter overlies the silt wherever the two are in contact. This contact may be gradational—that is, a 10- to 15-foot zone—but more often it is rela-

tively sharp. Where the contact is gradational, it consists of an increasing amount of organic material overlying a 2- to 4-foot zone of a green weathering surface of the silt. In this type of contact, the bedding of the muck and silt when visible is usually horizontal. Where the contact is sharp, the rude bedding of the silt, as well as its attitude as indicated by the ash layers, is usually horizontal, whereas the bedding of the muck locally parallels the contact, which roughly conforms with the slope of the valley and is at an angle of 5 to 15 degrees. The general relation of the muck and silt is shown in Figure 3.

ORIGIN

Toward the close of the period of stream aggradation, which is represented by the extensive auriferous gravel deposits, the silt deposition began. The present known extent and thickness of the silt indicate that its distribution was erratic but much greater immediately adjacent to the Tanana River than away from it. There must have been almost complete cessation of stream flow during the latter part of the silt deposition. The 140-foot exposure of silt at the mouth of Ester Creek was not entirely continuously deposited but was interrupted for short periods of time sufficient to develop a small amount of vegetation on exposed surfaces, as indicated by the carbonaceous remains in the ash layers. During these periods the climate was probably too cold or the period too short for the development of anything but a tundra-type of vegetation.

The most distinctive feature of the silt, as well as of the muck, is the uniformly fine-grained character of the mineral constituents; no coarse gradations have been found even immediately adjacent to the bedrock slopes. This character of widespread uniformity in texture is fortuitous, as in mining operations the silt and muck overburden is stripped from the gravel by water and is carried away in suspension without difficulty by ditches from 1 to 6 miles long, on as low grades as .15 to .3 per cent, which could only be effective on uniformly fine material. Only by an aeolian origin can this widespread uniformity in fineness be satisfactorily explained, and its greater abundance adjacent to the Tanana Valley gives an indication of the source.

The deposition of the silt or loess decreased as the climate became more humid and possibly warmer, and, at first, only mosses, which now grow at higher elevations, attained a foothold in the valleys. Streams began to erode the silt, but in protected places it weathered to depths of several feet. Vegetation continued to develop and extended up the stream valleys, and, with the coming of abundant plant life, animal life followed.

Some loess was probably still being deposited after vegetation had started, but in such small amounts that the plant life continued to flourish. As erosion continued, the loess was carried from the hillsides to the valleys and remained in place only where its thickness and extent were great or in protected localities. Much of the originally deposited loess has been transported out of the district, but, as the climate is semiarid, erosion has never been great in recent times, and, as vegetation became more abundant, more and more of it was held in the valleys. Stream erosion, sheet wash, creep, and slide have contributed to its redeposition in the valley bottoms. Deposition occurred so slowly that successive layers of vegetation were built up, and the surfaces were often permanent long enough to permit several feet of weathering and the accumulation of several feet of peat. The more frequent and thicker beds of peat at or near the surface indicate that erosion has continuously become slower, probably because of the aggradation of the local base level, the Tanana River, and the slowing down of erosion by the more abundant vegetation.

The thicknesses of muck now seen were not only built up slowly, but freezing occurred almost simultaneously. The winters were long and cold enough for the depth of annual freezing to exceed the summer thawing, so that as the depth of muck increased it always remained frozen. Because of its mode of origin the muck was saturated with water, giving it its present ice content of around 50 per cent. Vertebrate bones and skeletons with frozen flesh and skin still adhering are eloquent evidence of the almost simultaneous deposition and freezing of the muck.

The direct aeolian origin of the silt seems almost conclusive, but its contribution to the inorganic constituents of the muck is not known. Because of the similarity in size and type of minerals in both deposits, the bulk of the inorganic material of the muck is probably reworked silt. However, as creep, slide, and sheetwash were greatly effective in moving the silt from the higher slopes to the valley bottoms, some bedrock debris must also have been carried down and redeposited with the silt. In some localities—and here they always occur close to the valley walls—angular bedrock fragments in the muck are common. Comminution of bedrock is very effective in sub-Arctic climates and probably furnished some fine material.

As early as 1914, H. H. Bennett made a soil reconnaissance of the Fairbanks district for the Department of Agriculture and he classified the silt as a loessial soil.

Although no detailed work has been done on them, the writer, while examining mining districts from Dawson on the Upper Yukon to Mar-

shall on the Lower Yukon, has seen many deposits of muck and silt identical with those in the Fairbanks area, so that it seems highly probable that they have a similar origin.

OTHER THEORIES OF ORIGIN

GENERAL STATEMENT

Two alternative theories have been proposed—one, a lacustrine, and the other, the residual weathering of bedrock and the transfer and redeposition of the fine detrital material by solifluction processes.

LACUSTRINE

Because of the obvious terrestrial origin of the muck, a lacustrine origin could be applicable only to the silt. The uniform texture of the silt cannot be reconciled with a lacustrine origin, as, although material of this fineness would be deposited in certain portions of the basin, gradations in texture would occur—at the shore lines and where streams entered—and such gradations have never been found. In any large lake deposit, clay would form an appreciable part of the sediment. The silt contains an average of less than 3 per cent and it never occurs in lenses but is disseminated uniformly throughout the deposit. No evidence of lacustrine life has been found; instead, the fine carbonaceous layers in the ash suggest a terrestrial formation. If the silt represents a lacustrine deposit, there must have been an upper limit of deposition. A 1200-foot level has been suggested as the upper limit of deposition, because in the past it has been thought by some that the silt never occurred above that elevation. Although silt has not been found on the highest ridges, there is no sharp termination of it at any elevation and no topographic evidence of terraces representing its upper limit, although it is reasonable to assume that a few would have survived total dissection. Reconnaissance work on the Lower Yukon and Tanana rivers by the United States Geological Survey has failed to show evidence of a recent barrier sufficient to impound the water to a depth necessary to form the silt deposits at Fairbanks.

SOLIFLUCTION

An origin by solifluction processes for the silt and muck would mean that the bedrock has been the source of the inorganic material, and such a process would be accomplished through comminution of bedrock by weathering and the subsequent transfer of the rock debris to lower elevations by slide, creep, and sheetwash. Such a process must have played a part in the origin of the muck but probably was principally effective only in transporting the silt, which already mantled the hillsides, to the valley bottoms. In the origin of the silt, it is almost

impossible to conceive that the necessary selective sizing of material could have taken place. Bedrock weathering in sub-Arctic climates is dominantly mechanical disintegration with the production of only a small amount of clay but it has produced in this district much more clay than is now present in the silt. The absence of larger fragmental rock debris has compelled those who have favored a solifluction origin to advance an explanation for their elimination. Filtration through the abundant moss cover allowing the fines to pass through and retaining the coarse sizes has been suggested, but such a process of the magnitude necessary has never been observed. Even if such a process were effective, an inconsistency is present as the muck, which was formed when there was an abundant moss cover, occasionally contains large fragmental debris, whereas no such fragments have been found in the silt, which was deposited when there was little if any moss cover. At present, soil flows plainly of bedrock origin always contain bedrock fragments of varying size as well as an appreciable clay content, although the moss cover is abundant. Rude horizontal bedding and undisturbed horizontal ash layers in the silt in the Ester Creek area eliminate the possibility of its deposition by surface creep of the bedrock debris from the adjacent hillsides, as such structures should roughly conform with the slope of the terrain. Silt, up to 150 feet thick, often caps bedrock hills that are too far from the nearer higher hillsides for the latter to have contributed detrital material by solifluction processes. A similarity of the minerals in the silt and the bedrock has been suggested as evidence of a genetic relation. Such a similarity is true of the principal minerals—quartz, feldspar, and mica—but they are such common minerals that little weight can be given to them, and no work has been done on the more critical heavy minerals. The Birch Creek schist, which is the dominant bedrock in the Fairbanks district, has an areal extent of many thousands of square miles in the Tanana and Yukon valleys and on the north slope of the Alaska Range, so that almost any mode of origin could account for the common minerals present in the silt and muck.

LATE TERTIARY TO RECENT GEOLOGIC HISTORY

The Pleistocene and Recent history of the district is by no means entirely clear, but what is known is presented here, including the regional background of the silt and muck deposition.

During the latter part of the Tertiary, erosion was the dominant natural agent in the Fairbanks district as well as in a large part of the interior of Alaska, and the present late mature topography was being carved. A large part of the placer gold of the district occurs as a concentration on bedrock, a product of a long period of degradation, and

therefore principally a Tertiary deposit, whereas the associated gravel is younger. Erosion was not continuous, as drilling has indicated three interruptions to this cycle. None are visible on the present topography but are found only on the lower portions of streams that empty into the Tanana Valley, where they are now concealed by alluvium, silt, and muck. Topographically, there is some indication of a temporary base level at an elevation of 1000 to 1400 feet, but these breaks in the topography are results of rock structure. The first temporary base level was where the Tanana River flowed at an elevation of about 450 feet, and broad mature valleys were carved. Economically this period was most important, as it represents the longest continuous period of erosion, and therefore resulted in the largest gold concentration. Then the streams resumed their downcutting, and youthful valleys were carried down to an elevation of below sea level at Fairbanks; but, before mature valleys could be carved at this elevation, base level was raised by a tremendous alluvium fill, at least 500 feet thick, in the Tanana Valley at Fairbanks. Once more the streams renewed their erosion, and the present mature and late-mature bedrock valleys were carved. The great and apparently rapid changes in base level produced an extremely complicated erosional pattern of many drainage changes in the tributary streams close to the Tanana Valley. One very anomalous condition caused by the thick alluvial fill was the superimposition of a mature valley upon a youthful one. The great changes in base level in the Late Tertiary and Quaternary, and the present bedrock elevation of below sea level at Fairbanks, which is 1000 miles from the ocean, must be attributed to crustal warping. Warping has continued up to the present time, as on Lower Goldstream Creek there is a bedrock divide which can be explained only by warping, and in several localities there has been bedrock faulting of some of the gold-bearing channels.

The aeolian deposition of the silt seems indisputable and, if correct, is probably related to glaciation. The Yukon Basin is unglaciated, but, during the Pleistocene, glaciers extended out from the Alaska Range, and conditions for loess deposition in the Fairbanks district were ideal. Numerous glacial streams, heavily laden with glacial flour, deposited it on the river bars, whence prevailing off-glacier winds carried it away and redeposited it. Such conditions—a common result of glaciation and even now occurring on the Tanana Flats and elsewhere in Alaska (Tuck, 1938)—must have been much more ideal during the height of glaciation when vegetation was less abundant, silt-laden streams larger and more numerous, and prevailing winds stronger. It is significant that the typical muck and silt deposits are most extensive in the unglaciated

part of Alaska bordering the major streams, and none of the type that has been described here are found in the glaciated portions.

In a large part of the unglaciated portion of Alaska the subsoil is perpetually frozen. The conformity of the area of perpetually frozen subsoil with the unglaciated area may be more than a coincidence; the glaciated area was probably insulated from the prevailing cold of the glacial period by its ice blanket, whereas, in the unglaciated area, unprotected by any ice cover, both the alluvium and the bedrock were frozen to depths of a few feet to 300 feet.

In the Fairbanks district, the muck always overlies frozen gravel, whereas the silt overlies unfrozen gravel (Fig. 3). The silt is plainly older than the muck, and in all known occurrences the gravel underlying the silt is older than the adjacent frozen gravel. If the perpetually frozen condition of the gravel were caused by its lying exposed during the height of glaciation, then the unfrozen gravel must have been protected even at that time by silt, so that the silt was deposited during the advancing stage of glaciation, and the muck represents a retreating stage continuing up to the present. The present annual temperature of the Yukon Basin is slightly below freezing; during the Pleistocene, with large ice sheets in adjacent districts, the climate must have been much colder. It seems reasonable to assume that during a large part of the glacial period the Fairbanks district, as well as a large part of the interior of Alaska, lay bare, windswept, and with no vegetation except possibly a tundra type. The deposition of the silt and the absence of vegetation in the advancing stage of glaciation suggest slight rainfall. A dominant factor in causing glaciation in Alaska may then have been decrease in annual temperature. The great alluvial fill in the Tanana Valley indicates that the necessary decrease in temperature could have been due to elevation of landmass, subsequent depression causing a recession of the glaciers and the alluvial filling of the valley.

A study of growth rings to determine the age of trees buried in the muck on Engineer Creek in the Fairbanks district has been made by J. L. Giddings (oral communication). He has determined that trees buried 30 feet below the surface, and in place, date from 1300 to 1450 A.D. Numerous scattered bones of species now extinct, such as mammoth and mastodon, are found near the same horizon. It is questionable whether these animals existed in this district only 500 to 600 years ago, as the bones may have been buried in and derived from an older muck deposit. However, it is evident that these animals existed in this area almost within historical times and under geologic and climatic conditions similar to those found in other parts of the world.

CONCLUSIONS

The origin of the muck-silt overburden which blankets the extensive gold-bearing gravel deposits in the Fairbanks district has long been obscure because of poor exposures and the complications of sub-Arctic solifluction processes. It now seems certain that the silt is of aeolian origin, while the inorganic material of the muck consists principally of wind-blown particles that have been reworked by creep, sheetwash, and stream action. A large part of the interior of Alaska is unglaciated; the deep muck-silt deposits occur in this unglaciated area and are probably a result of Pleistocene glaciation in nearby localities, as conditions were then ideal for aeolian deposition.

WORKS TO WHICH REFERENCE IS MADE

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Tuck, Ralph (1938) *The loess of the Matanuska Valley, Alaska,* Jour. Geol., vol. 46, no. 4, p. 647-653.

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TYPE LOCALITY OF THE CRETACEOUS CHICO FORMATION

BY J. A. TAFF, G. D. HANNA, AND C. M. CROSS

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ABSTRACT

Inadequate reliable information on the Chico formation of the California Cretaceous led the writers to re-examine the type locality. Evidence is presented to show that Chico Creek must be considered to be the type locality, and an historical review of previous work is given. The Cretaceous rocks exposed in Chico Creek comprise a series of clastics 2000 feet in thickness, resting unconformably on the older Sierra Nevada metamorphics and covered at the top by Tertiary volcanics. The fauna of the Chico beds exposed in Chico Creek is abundant, but only meager evidence exists to correlate it with other Cretaceous localities in the central valley of California. Correlation with other more removed sections is difficult, and a conclusive tie into the European section cannot be made at this time. Typical fossils found in Chico Creek are discussed, and neotypes are illustrated for certain important species, the originals of which have been lost (*Ammonites chicoensis* Trask and *Baculites chicoensis* Trask). All figured specimens are deposited in the California Academy of Sciences.

INTRODUCTION AND ACKNOWLEDGMENTS

In conducting certain studies of the Cretaceous along the west side of the San Joaquin Valley in California, it became impossible to reconcile some of the authors' observations with published information regarding the Chico. The formation (or group) itself did not appear to have received adequate definition, and as time went on it became increasingly apparent that further investigation of a typical section would be necessary because of the widespread misuse of the name Chico in geologic

literature. Several localities were considered for this, and, after study of the published information regarding all of them, it appeared that these researches should be made in Chico Creek, the type locality of the formation. Therefore, in May and October 1934 and February 1935, the writers visited the northern part of the Sacramento Valley; on each occasion Chico Creek and other related areas in the district were examined. It has been concluded that Chico Creek must be considered the type locality in any discussion of the rocks which bear that name.

Formal designation was definitely made by Stanton (1896, p. 1013) who stated that

". . . the rocks . . . [there] are now known to be Cretaceous, and this must be regarded as the typical locality of the Chico group to which it gave its name."

The authors wish to acknowledge the assistance and helpful criticism given by L. G. Hertlein, S. W. Muller, and F. M. Anderson in connection with the preparation of the paper.

PREVIOUS WORK

The name "Chico Group" was proposed by Gabb (1869, p. xiii, xiv, 129, footnote) in a paper read before the National Academy of Sciences, Northampton, Massachusetts, in August 1868. In December of that year, or early in 1869, he published the name in the following words:

"I shall, further, in this paper, use the term Martinez Group, for the upper portion of 'Division A' of the California Reports; the Chico Group, for those beds of which Chico Creek, Pence's Ranch, and Tuscan Springs, are typical localities."

Earlier in 1868 (probably May or June) he referred to a species, "*Pugnellus (Gymnarus) manubriatus* Gabb," as having come from the "Chico Group," but this can in no way be considered a definition of a formation. At that citation he stated the age to be "lower chalk or upper green sand."

Although the name of the formation must date from 1868 or 1869, the first fossils from it were described 12 years earlier. In 1856 Trask described *Ammonites chicoensis* and *Baculites chicoensis* from Chico Creek; he also illustrated members of several genera of gastropods and bivalves whose nature caused him to believe that the age of the rocks was Tertiary.

Trask's attention was called to the outcrops on Chico Creek by Dr. Andrew Randall, whose name does not appear otherwise in connection therewith. The collection was made during the winter of 1855-1856 and was later deposited in the California Academy of Sciences; it was made available for the use of the members of the Pacific Railroad Survey and the State Geological Survey.

Blake (1857) was given a piece of the fossiliferous Chico Creek rock by Trask and from it he reported "*Nucula divaricata* Conrad, *Mactra albaria* Conrad, *Tellina, Natica, Dentalium* and *Baculites*." These fossils led him to believe the age to be Miocene. According to Newberry (1856) the first two were described from the Miocene of Astoria, Oregon, and were determined by Conrad.

Newberry (1856) stopped at Bidwell's place on Chico Creek—where the town of Chico now stands—and found fragments of the fossiliferous sandstone washed down from exposures to the east. In addition to the forms mentioned by Blake he listed "*Fusus*" and "*Turritella*" and added in a footnote that Meek had found "*Inoceramus*" in the rock. Both Newberry and Meek were inclined to regard the age as Cretaceous.

On October 11, 1862, a party from the California Geological Survey, composed of Brewer (1930), Rémond, and Schmidt, visited Chico Creek to collect fossils. Nothing is said of their observations by Whitney in his volume 1 of the Survey Reports, but presumably the fossils were used by Gabb in his work on the paleontology. The only definite available account the writers found is by Brewer (1930, p. 339), and he merely states:

> "The volcanic deposits here were about eight hundred feet thick, lava and ashes interstratified. The stream has cut entirely through into the sandstones beneath, which teem with shells. They are fossils, but are apparently as fresh as if left on the beach but a few years ago—only imbedded in sandstone. Large masses seemed half made of shells."

Descriptions and notes on 19 species were recorded by Gabb in 1864, and some changes in names appeared 4 years later. Whitney (1865, p. 208, 209) apparently based his brief notes on the locality upon the records of Trask, Brewer, and Gabb.

About 40 years ago Diller, Stanton, and Turner, of the United States Geological Survey, made extensive studies of the geology of the Sacramento Valley, including the area of the type Chico. The first report is by Diller (1889, p. 407-410). He collected a few fossils "15 miles east of Chico," and these were determined by White as:

> "*Trigonia Evansana* Meek, *Venus varians* Gabb, *Fulguraria Gabbi, Cinulia obliqua* Gabb, *Turritella uvasana* Con., *Baculites chicoensis* Trask, *Aporrhais falciformis* Gabb, *Ammonites Newberryanus* Meek, *Ammonites chicoensis* Trask, Gastropod sp."

White (1889) described numerous fossils from several localities supposed to be Chico in age, but only one—"*Vasculum obliquum*"—is definitely recorded from Chico Creek.

Apparently the fossils collected by Diller were later studied by Stanton (1896) and the results appear to have been used by Turner (1896). The

latter visited Chico Creek in 1885 and described a collecting locality which the present writers have determined to be the uppermost exposure of the Chico section there. It is noted herein as C. A. S. Loc. No. 28171. The fossils he collected were determined by Stanton and were listed in 1894.

In discussing the relationship of the Cretaceous of the Santa Ana Mountains, Packard (1916, p. 153-155) gave some notes on the type Chico of Chico Creek, based largely upon the work of C. K. Studley. Collections made by the latter enabled Packard to give a list of 24 species which were also found in the Santa Ana Mountains. An additional two, "*Pecten operculiformis* and *Nemodon vancouverensis*," were included on page 153, but it is possible that this was unintentional; no other mention of them from Chico Creek has been found.

In a later paper, Packard (1916, p. 298-300) gave detailed notes on the three species of *Spisula* which he had listed.

The investigations of Studley (1912) were made under the direction of J. C. Merriam and were incorporated in a thesis presented to the faculty of the University of California. A copy of this manuscript in the University Library has been consulted, and the information gained was found decidedly helpful in some of the field work. Studley's check list contained the names of 75 species of fossils from four localities on Chico Creek. These include all the previous records he was able to find in the literature, as well as his own collections. He described several species as new, but the paper is without illustrations, except for some geologic sections.

Stewart, in 1927 and 1930, published a revision of the type material (gastropods and lamellibranchs) which was used by Gabb in preparing the volumes of the Geological Survey of California and which was still to be found in the Academy of Natural Sciences of Philadelphia. The discussions of genera and species are usually comprehensive. It was shown that Chico Creek is the type locality for only four species, and these Stewart (1930, p. 9, 10) has recorded as *Anchura falciformis* (Gabb), *Turritella chicoensis* Gabb, *Flaventia ? lens* (Gabb) and "*Dosinia*" *inflata* Gabb.

Finally, Popenoe published a study of the Cretaceous of the Santa Ana Mountains, California, in 1937, in which he made incidental reference to two species which also occur at the type locality of the Chico. These are *Trinacria cor* Popenoe and *Turritella chicoensis* Gabb.

His paper contains the results of careful exposure of the hinges of many bivalve mollusks, and consequently he was able to place most of the species in genera with greater assurance than was previously possible.

This necessarily involved the shifting of some well-known type Chico species, particularly the following:

Older name	Popenoe's determination
Dosinia inflata Gabb	*Tenea inflata* (Gabb)
Meretrix lens Gabb	*Flaventia lens* (Gabb)
Meretrix nitida Gabb	*Calva nitida* (Gabb)
Venus varians Gabb	*Calva varians* (Gabb)
Mactra (or *Spisula*) *ashburneri* Gabb	*Cymbophora ashburneri* (Gabb)

In a late paper Popenoe (1940, p. 163) showed that the name *"Calva"* was a synonym of *Trigonocallista* Rennie (Rennie, 1930).

There is an extensive literature on the geological relationships of the various divisions of the Cretaceous of northern California, most of which contains incidental reference to the "Chico" as a formation or group name. None of the papers, however, other than those mentioned by the present writers, are the result of actual work in Chico Creek or renewed investigations of the fossils from there. They are essentially the results of field work elsewhere in the area with conclusions as to Chico Creek drawn from previous literature. Some of the more important of these papers are by Diller (1893), Stanton (1893), Diller and Stanton (1894), and Bryan (1923).

GEOLOGY

The type locality of the Chico is in the Canyon of Chico Creek, 8 to 16 miles northeast of the city of Chico, Butte County, California. Chico Creek, in its descent through the foothills of northern Sierra Nevada Mountains to the Sacramento Valley, has cut a gorge 1000 to 1500 feet deep through late Tertiary lava flows and tuff beds of the Tuscan formation, exposing the Chico, upper Cretaceous strata (Figure 1). In 1933, C. A. Anderson published the results of extensive studies of volcanic rocks of an area in northern California which includes Chico Creek. His Plate 17, fig. *c*, and Plate 18, fig. *a* show views of these materials in this creek, and incidental reference to them is made at several places in the text. Figure 3 [8] opposite page 218 is a columnar section of the Tuscan formation in Chico Creek.

Other areas of upper Cretaceous rocks are exposed on the eastern border of the Sacramento Valley from the north end opposite Redding, Shasta County, to the vicinity of Folsom, east of Sacramento. These areas, like that of the type locality in Chico Canyon, are exposed by incised streams, which have removed unconformable Tertiary volcanic or sedimentary beds. These several areas of upper Cretaceous, like that of the Chico type locality, rest unconformably upon older Mesozoic or Paleozoic rock and have the same general trend and attitude of low

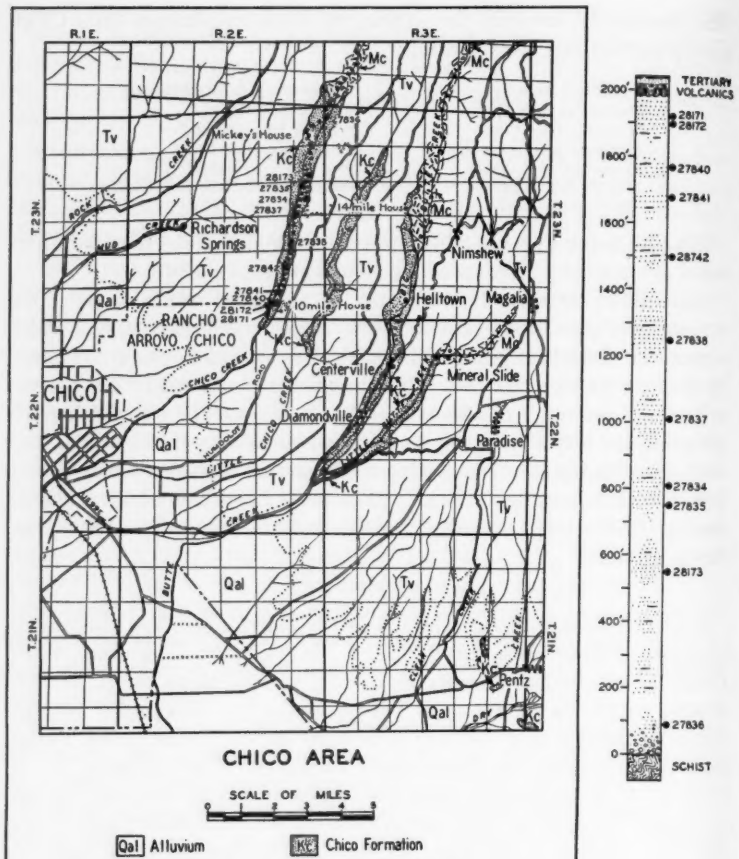

FIGURE 1.—*Portion of Chico Quadrangle and columnar section*
Showing location and extent of the geologic units in the vicinity of Chico Creek, Butte County, California. After United States Geological Survey topographic map.

dip westward beneath the valley. Paleontological evidence on hand, however, is not sufficient to place most of them in their stratigraphic position with respect to the Chico in the type locality.

The areas exposed in Chico Canyon are more extensive and more accessible than in any other locality in the vicinity of Chico; hence its

choice by the earlier explorers as the type locality of the Chico formation was fortunate. The strata there are of bluish-gray arkosic sandstone, weathering to olive drab and greenish buff, interbedded with a few thin layers of sandy shale and occasional thin layers of carbonaceous shale. At and near the base in Chico Canyon, the sandstone contains beds of conglomerate of dense and smoothly rounded quartzite, quartz, schist, and granitic rock with the common sand matrix.

The total thickness of the Chico formation at the type locality is unknown. The part exposed, however, from the base to the top of the exposed section is, by estimate, 2000 feet thick (Fig. 1).

The Chico beds, resting unconformably upon steeply inclined metamorphosed schistose shales and sandstone, dip monoclinally at an average of 10 degrees approximately S. 70° W. and, in turn, are overlain unconformably by stratified tuffs and lava beds of the Tuscan formation which dip southwesterly approximately 2 degrees. The Tuscan formation overlaps the bevelled edge of the Cretaceous and extends toward the northeast upon the older schist. Toward the southwest the same tuff and lava descend to the level of the streams overlapping the Cretaceous near the Sacramento Valley border. The Chico exposures, therefore, form inliers along the canyon walls and in the banks and bed of Chico Creek.

The overlying stratified tuff and lava beds, 500 to 700 feet thick, in the upper walls of the canyon, break down forming a mantle of talus and soil over the edges of the softer Chico strata so that fresh exposures occur only at intervals in the slopes of the canyon and along the banks and bed of the stream.

A search of the type locality from the base to the top of the Chico exposures revealed 12 locations distributed through the section which yielded fairly abundant fossils. The lowest exposed fossil horizon is approximately 100 feet above the base, and the highest is about the same distance below the top. The location of exposures of Chico strata at the type locality where fossils have been collected is shown on Figure 1. A columnar section of the strata exposed in Chico Canyon showing the stratigraphic positions of the fossil collections is drawn on the border of the map.

CORRELATION

The exposures on Chico Creek were found to form a unit without any stratigraphic break, and, since certain zones are highly fossiliferous, it would seem that correlation with other localities would be relatively simple. However, the contrary has proved to be the case. Either the assemblage of species represented there occupied peculiar environmental

conditions and was, therefore, somewhat local in nature, or equivalent strata are not well known in other parts of California. It is obvious that the waters during the accumulation of these strata were purely marine in character and it is equally obvious that at least most of the sediment was derived from the subsiding land mass to the east.

Unfortunately, the volcanic activity of the Tertiary covered much of the foothill country of the eastern part of the Sacramento Valley, and it is only where erosion has had a considerable effect that the underlying rocks may be seen. Thus only a few of the larger streams in addition to Chico Creek have cut through the mantle of volcanics. This is true of Butte Creek not far to the south and at Pentz' (older spellings "Penz" or "Pence") Ranch, somewhat farther south. At the latter place frontal erosion of the flow rocks has exposed Cretaceous rocks, and at all these nearby localities it has not been difficult to determine the fossil content of the sediment as Chico.

Northward from Chico Creek there is no such certainty of position. Characteristic fossils which would be expected to be present have not been found, and consequently the geological relationships remain obscure.

From Redding southward, along the west side of the Valley, conditions grow more and more obscure. The types of sediment have changed from those of Chico Creek, and fossils in those parts of the section where equivalent strata might be expected are nearly nonexistent. A good assemblage of fossils comparable to those of the Chico Creek localities has not been found in place anywhere along the west side of the Sacramento-San Joaquin Valley border, a distance of about 400 miles, with many magnificently exposed sections.

It might be expected that a well drilled somewhere out on the valley floor would furnish evidence which would help to extend the Chico strata westward. Those which have been drilled have either been unfortunately located or for some other reason have failed to give information of the slightest importance in this problem. In a few cases microfossils have appeared in the cores of some of the wells but they are neither abundant nor comparable with the fauna of the type locality. The writers were unable to find any microfaunas in the samples taken in Chico Creek.

One very peculiar circumstance connected with the west side sediments may help to place the Chico in its proper position in the geologic column. Some conglomerates are widely distributed on that side, and, while a single bed probably does not extend throughout the area, outcrops of some can be traced for many miles. In a few favorable places one of these conglomerates can be accurately located in the section by overlying or underlying beds carrying some characteristic fossils. Within these conglomerates at many widely distributed localities fossiliferous

boulders have been found. While no ammonites have thus far been discovered, pelecypods and gastropods are abundant in the boulders. The character of the sediments in the boulders and the fossils themselves, both as to abundance and specific identity, strongly suggest that the boulders have been derived from beds equivalent to those in Chico Creek. In many cases it would be extremely difficult to justify calling them by any other formation name. Thus these presumably Chico boulders have been found in Panoche Hills, Fresno County, in a conglomerate 9000 feet or more below the top of the Cretaceous section there exposed, and similar ones are common southward, at least to Orchard Peak, Kings County. If these boulders are Chico, as they seem to be, then that formation was in existence long enough to become thoroughly solidified and calcified, raised above sea level, and eroded when the conglomerates were deposited. Obviously any subsequently deposited sediment would necessarily be post-Chico.

Except for these boulders, no satisfactory correlation has been made between any of the sediments found on the west side of the San Joaquin Valley and the type Chico beds in Chico Creek.

Passing farther afield there is good fossil evidence, as Packard and Popenoe have shown, to prove the existence of strata equivalent to the Chico formation, or nearly so in southern California, particularly in the Santa Ana Mountains. Some of the ammonites Anderson (1905) described from Oregon bear at least generic resemblance to those from Chico Creek. This would indicate that the strata were similar in age and perhaps equivalent. A careful world-wide study of the literature on ammonites would probably yield some information of indirect value in the proper dating of the Chico.

DISCUSSION OF THE FOSSILS

In a few favorable localities in Chico Creek, fossils are abundant in calcareous concretionary masses in the sandstones. Elsewhere they are usually scarce, even as impressions. These aggregations of cemented sands are very hard and tough so that the extraction of the fossils, which are somewhat chalky and fragile, is, to some extent, uncertain and tedious. In spite of these difficulties, a fairly large collection was made during this investigation and has been deposited at the California Academy of Sciences. It is not practicable at this time to present a monograph of the fauna found in the section. Such a study would require an extended period, and the writers do not feel competent to undertake such a specialized task. However, it is believed that the importance of the locality warrants the illustration of some of the most common and representative

forms so that they can be searched for elsewhere in attempting correlations with this period of Cretaceous deposition. In making the selection shown, the writers have been guided primarily by the presence in this fauna of several peculiar and rather distinctive types, none of which are known to have a very extensive stratigraphic range but all of which are believed to have a fairly wide distribution along the west coast of North America. Exact correlations with other localities, both to the north and to the south, must be left to those who may make careful comparative studies. The species selected have been assigned the names which are believed to represent the latest investigations into Cretaceous taxonomy. The assistance of F. M. Anderson in this matter is gratefully acknowledged.

Ammonites chicoensis Trask

(Plate 1, figures 1, 2)

Ammonites chicoensis TRASK, Calif. Acad. Nat. Sci., Pr., 1856, vol. 1, p. 92-93, pl. 2.

The history of this name and the circumstances attending the discovery of the species were given above. The original specimen was evidently immature, and the figure and description are not adequate for specific identification. No subsequent author has mentioned having seen the specimen. It is practically certain to have been lost in 1906 when the collection of the California Academy of Sciences was nearly all destroyed by fire. Gabb (1864, p. 68, pl. 13, fig. 17) described and illustrated what he took to be *Ammonites chicoensis*. Several localities were listed, but he did not give the station for the one shown on Plate 13. Merriam, in 1895, did not find the specimen when he listed the types of the Geological Survey which had been deposited in the University of California. Inquiry at several eastern museums has failed in locating it. If it should be found it will automatically become the holotype of *"Schloenbachia" gabbi* of Anderson (1905, p. 63) who was unable to identify Trask's species definitely in the material at the University of California in 1905. It is interesting to note that this particular specimen of Gabb's interpretation of *chicoensis* was used by the Geological Survey of California to make the embossing stamp used to impress the gold emblem on the binding of volume 1 of the *Palaeontology* and what appears to be the first issue of volume 2; a different emblem and binding were used on the second issue of the latter.

In view of the uncertainty as to which ammonite should be assigned to the name *chicoensis,* the rules of nomenclature in force at present offer a means of stabilization. This is to select a neotype, which in all essential respects can be referred to with the same confidence that is usually placed in a holotype. This the writers have done, with the advice and assistance of their associates, from a station on Chico Creek (Loc. 27838 C. A. S.) 3.1 miles south of bridge at Frank Mickey's ranch, or 3.6 miles from "10 mile house" (U. S. Geological Survey topographic sheet) on the Humboldt Road, east of Chico, Butte County, California, May 3, 1934, and the specimen is hereby designated neotype of *Ammonites chicoensis* Trask, no. 5785 Calif. Acad. Sci. coll. F. M. Anderson is authority for the statement that it probably belongs to the genus *Mortoniceras,* but until comparatively recent times it would have been called *Schloenbachia.*

Baculites chicoensis Trask

(Plate 1, figures 3, 4)

Baculites chicoensis TRASK, Calif. Acad. Nat. Sci., Pr., 1856, vol. 1, p. 92-93, pl. 2.

The status of this name is nearly the same as that of *Ammonites chicoensis*. The original specimen is lost, and doubt exists as to subsequently identified material. Gabb (1864, p. 80, Pl. 14, figs. 27, 29; Pl. 17, fig. 27) illustrated several things which he called by that name. Of these, the one represented by his Figure 29 was stated by Merriam, in 1895, to be in the University of California collections, but this is a transversely ribbed form, obviously not the one Trask had. In order to clear the matter for subsequent investigators we have selected from Loc. 27838 (C. A. S.) (Chico Creek, 3.1 miles south of bridge at Frank Mickey's ranch, 3.6 miles from "10 mile house" on the Humboldt Road east of Chico, Butte County, California) two specimens to serve as neosyntypes of the species. They bear nos. 5786 and 5787, Calif. Acad. Sci. coll. and are nearly smooth, with two parallel grooves on the ventral side forming, between, a rounded ventral keel-ridge. The species is very common at the type locality and is the only *Baculites* detected in the collections made in connection with this study.

Cucullaea truncata Gabb

(Plate 1, figures 5, 6)

Cucullaea truncata GABB, Calif. Geol. Survey, vol. 1, Palaeont., 1864, p. 196, pl. 25, fig. 182.

Cucullaea (*Cyphoxis*) *truncata* Gabb, STEWART, Philadelphia Acad. Nat. Sci., 1930, Spec. Publ. 3, p. 75, pl. 2, fig. 7.

This robust form is abundant and well preserved in several of the Chico Creek localities. Radial ribs are reduced to fine irregular lines. The species is representative of a group which, during this part of the Cretaceous, was very widely distributed; consequently, it has been cited in many places in geological literature.

Trigonocallista varians (Gabb)

(Plate 2, figures 10, 11)

Venus (*Mercenaria?*) *varians* GABB, Calif. Geol. Survey, 1864, vol. 1, Palaeont., p. 161, pl. 23, figs. 140, 140a, "141." "Martinez and Pence's ranch, Butte County."

Aphrodina varians (Gabb) STEWART, Philadelphia Acad. Nat. Sci., Spec. Publ. 3, 1930, pl. 249, pl. 6, fig. 6; [same specimens as Gabb's fig. 140a].

Calva varians (Gabb) POPENOE, Jour. Paleont., vol. 11, no. 5, 1937, p. 394.

Trigonocallista varians (Gabb) POPENOE, Jour. Paleont., vol. 14, no. 2, 1940, p. 163.

Gabb illustrated three specimens, one of which (fig. 140a) was chosen as the lectotype by Stewart; it apparently came from near Martinez. Numerous specimens from Chico Creek seem to be the same. Gabb's figure 141 of a specimen from Pentz is much higher, and it is suggested that it may be a fully grown *Trigonocallista nitida*. *"Venus" varians* has been referred to very often in west coast Cretaceous literature, and while, in some cases, the specific identification may have been doubtful, nevertheless, shells of this form are very common and readily recognized by the prominent, widely spaced, concentric grooves.

Trigonocallista nitida (Gabb)

(Plate 2, figures 3, 4)

Meretrix nitida GABB, Calif. Geol. Survey, vol. 1, Palaeont., 1864, p. 165, pl. 23, figs. 145, 146. "Near Martinez; *Arestimba cañon;* Chico and Cow Creeks."

Aphrodina nitida (Gabb) STEWART, Philadelphia Acad. Nat. Sci., Spec. Publ. 3, 1930, p. 250, pl. 5, fig. 10, pl. 6, fig. 9.

Calva nitida (Gabb) POPENOE, Jour. Paleont., vol. 11, no. 5, 1937, p. 394.

Trigonocallista nitida (Gabb) POPENOE, Jour. Paleont., vol. 14, no. 2, 1940, p. 163.

The original specimens from Martinez described and figured by Gabb seem to have been lost, but Stewart tentatively selected two from the Geological Survey collections now in the Philadelphia Academy for figuring and one for a lectotype; presumably they came from the same locality. The species is common in Chico Creek and seems to be constantly separable from *T. varians* (Gabb), by being shorter, higher, and more inflated. Both species are marked by prominent concentric grooves, widely spaced, representing rest periods in growth. Both species have been assigned to numerous genera. Popenoe excavated the hinge line and found a new genus name was needed for the group and selected *"Calva"(-Trigonocallista)* for the purpose. The writers' best specimen has both valves attached and shows the external characters very well.

Tenea inflata (Gabb)

(Plate 2, figures 1, 2)

Dosinia inflata GABB, Calif. Geol. Survey, vol. 1, Palaeont., 1864, p. 168, pl. 23, fig. 149.—STEWART, Philadelphia Acad. Nat. Sci., Spec. Publ. 3, 1930, p. 231.

Tenea inflata (Gabb) POPENOE, Jour. Paleont., vol. 11, no. 5, 1937, p. 391, pl. 48, figs. 3, 5.

Gabb's figure of this species is not very good, and the exterior of another specimen from Chico Creek, the type locality, has not since been presented. Stewart indicated that the holotype was no. 31445 of the University of California collection, Merriam (1895) having previously indicated that it was located there. Popenoe illustrated the hinge from material collected on Butte Creek, not far from the type locality.

Auchura falciformis (Gabb)

(Plate 2, figures 7–9)

Aporrhais falciformis GABB, Calif. Geol. Survey, vol. 1, Palaeont., 1864, p. 127, pl. 20, fig. 83.

Auchura falciformis (Gabb) Am. Jour. Conchol., vol. 4, 1868, p. 145, pl. 14, fig. 14.—STEWART, Philadelphia Acad. Nat. Sci., Pr., vol. 78, 1926 [1927], p. 360, pl. 22, fig. 9. Lectotype selected "probably from Chico Creek."

This strongly ornamented species is very common in Chico Creek exposures and has been reported from numerous other localities. It is probably confined to the stage of the Cretaceous to which the Chico formation belongs. Stewart, in his study of Gabb's type material from California, selected and figured a lectotype and pointed out that the original illustrations were composite drawings. This lectotype was believed to have come from Chico Creek. The ornamentation of the shell makes

extraction of suitable material for illustration difficult, but it is believed that the specimens shown herewith display the essential characters better than any which have been previously available.

Oligoptycha obliqua (Gabb)

(Plate 2, figures 5, 6)

Cinulia obliqua GABB, Calif. Geol. Survey, vol. 1, Palaeont., 1864, p. 111, pl. 19, fig. 64, a-c. Tuscan Springs; Martinez; Pence's Ranch, above Oroville; Chico Creek; Texas Flat, Placer County (Division A). Also, at Cottonwood Creek; and in the Siskiyou Mountains, Siskiyou County.

Oligoptycha obliqua (Gabb) STEWART, Philadelphia Acad. Nat. Sci., Pr., vol. 78, 1926 [1927], p. 436, pl. 24, fig. 14. Lectotype selected, "probably from Tuscan Springs."

This small gastropod is very common in the collections from Chico Creek and has been widely reported from other localities. Stewart has given references to pertinent literature in his study of Gabb's type material and from it he selected and figured a lectotype. The exact locality from which it was obtained is not certain, but he thought it came from Tuscan Springs in Tehama County. The species is believed to be confined to that part of the Cretaceous which includes the Chico formation exposed on Chico Creek.

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Presented before the Cordilleran Section of the Geological Society, April 13, 1935.

EXPLANATION OF PLATES

PLATE 1.—FOSSILS FROM THE TYPE LOCALITY OF THE CHICO FORMATION

Figure

1, 2. *Ammonites chicoensis Trask.* Neotype no. 5785, Calif. Acad. Sci. from loc. 27838 (C. A. S.), Chico Creek, Butte County, California, 3.6 miles from "10 Mile House" on Humboldt Road (U. S. G. S. Topog. map). Chico Cretaceous. Greatest diameter, 79.6 mm.; least diameter, 60.5 mm.; greatest thickness, 26.0 mm. (p. 1320.)

3, 4. *Baculites chicoensis* Trask. Neosyntypes nos. (fig. 3) and 5786 (fig. 4) 5787. Calif. Acad. Sci., from loc. 27838 (C. A. S.), Chico Creek, Butte County, California, 3.6 miles from "10 Mile House" on Humboldt Road (U. S. G. S. Topog. map). Chico Cretaceous. (3) Length of fragment, 78.8 mm.; greatest diameter, [of fragment], 18.0 mm.; (4) length of fragment, edge view, 77.7 mm.; least diameter, 11.0 mm. (p. 1321.)

5, 6. *Cucullaea truncata* Gabb. Hypotype no. 5791, Calif. Acad. Sci., from loc. 28172 (C. A. S.) Chico Creek, Butte County, California, at bluff on creek due west of "10 Mile House" on Humboldt Road (U. S. G. S. Topog. map). Chico Cretaceous. Length, 65.5 mm.; height, 59.6 mm.; thickness, 36.2 mm. (p. 1321.)

FOSSILS FROM TYPE LOCALITY OF CHICO FORMATION

FOSSILS FROM TYPE LOCALITY OF CHICO FORMATION

PLATE 2.—FOSSILS FROM THE TYPE LOCALITY OF THE CHICO FORMATION

Figure

1, 2. *Tenea inflata* (Gabb). Hypotype, no. 5790. Calif. Acad. Sci., from loc. 27838 (C. A. S.), 3.6 miles from Chico Creek, Butte County, California, "10 Mile House" on Humboldt Road (U. S. G. S. Topog. map). Chico, Cretaceous. Length, 28.9 mm.; height, 31.0 mm.; thickness, 14.6 mm. (p. 1322.)

3, 4. *Trigonocallista nitida* (Gabb). Hypotype, no. 5789, Calif. Acad. Sci. from loc. 27836 (C. A. S.) Chico Creek, Butte County, California, west side of canyon above conglomerate about 1.2 miles northeast of Frank Mickey ranch. Chico, Cretaceous. Length, 27.0 mm.; height, 25.8 mm.; thickness (both valves), 17.0 mm. (p. 1322.)

5, 6. *Oligoptycha obliqua* (Gabb). Hypotypes, no. 5792, 5793, Calif. Acad. Sci., from loc. 27838 (C. A. S.), Chico Creek, Butte County, California. 3.6 miles from "10 Mile House" on Humboldt Road (U. S. G. S. Topog. map). Chico, Cretaceous. (5) no. 5792; height, 12.3 mm.; diameter, 11.4 mm.; (6) no. 5793; height, 9.6 mm.; diameter, 10.0 mm. (p. 1323.)

7-9. *Auchura falciformis* (Gabb). Hypotypes, nos. 5794, 5794A, 5794B, Calif. Acad. Sci., from loc. 27838 (C. A. S.), Chico Creek, Butte County, California, 3.6 miles from "10 Mile House" on Humboldt Road (U. S. G. S. Topog. map). Chico, Cretaceous. (7) no. 5794A; length, 38.7 mm.; diameter, 14.0 mm.; (8) no. 5794B; length 26.0 mm.; diameter, 11.4 mm.; (9) no. 5794, length 47.9 mm.; diameter, including wing, 49.8 mm. (p. 1322.)

10, 11. *Trigonocallista varians* (Gabb). Hypotype, no. 5788 Calif. Acad. Sci., from loc. 27838 (C. A. S.), Chico Creek, Butte County, California, 3.6 miles from "10 Mile House" on Humboldt Road (U. S. G. S. Topog. map). Chico, Cretaceous. Length, 48.8 mm.; height, 43.5 mm.; thickness, 20.0 mm. (p. 1321.)

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 51, PP. 1329-1336, 6 FIGS. SEPTEMBER 1, 1940

GLACIATION OF THE GRAYS LAKE, ILLINOIS, QUADRANGLE

BY WILLIAM E. POWERS AND GEORGE E. EKBLAW

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ABSTRACT

The Grays Lake quadrangle in northern Illinois lies within the Valparaiso (Wisconsin) moraine. The late glacial history of this region embraced (1) deposition of moraines of the Tazewell substage, (2) deposition in late Marseilles time of an outwash plain inside the Tazewell moraines, (3) a post-Tazewell erosion interval in which the Fox River valley and tributaries were cut to present depth, and (4) glacial advance in the Cary substage, with deposition of the Minooka, Valparaiso, Tinley, and Lake Border moraines.

The Grays Lake quadrangle contains the eastern edge of the late Marseilles gravel plain, deeply trenched and later mantled by Valparaiso drift. A lowland east of this plain, probably a fosse, is now partly filled with moraine and low outwash of Valparaiso age. The Valparaiso till sheet is thin and rests on older drift. Widely distributed throughout the Grays Lake area below the 830-foot level are deposits of thinly laminated lacustrine clay and silt. Some of the silt rests on Valparaiso moraine and originated in shallow postglacial lakes, but much of it is marked by morainal topography, and some is overlain by Valparaiso till. Thus is indicated a pre-Valparaiso lake formed during the deposition of the Minooka moraine, first of the Cary substage. The Minooka glacier crossed Fox River valley at Elgin, Illinois, forcing it into a new 820-foot channel farther west. From Elgin the Minooka moraine trends northeastward and, although buried by Valparaiso drift, lies near the east margin of the Grays Lake quadrangle all of which west of the Minooka moraine was flooded to the 820- to 830-foot level. Silt and clay deposits of the lake thus formed were later overridden by the glacier in Valparaiso time. To this lake the name "Glacial Lake Wauconda" is given.

GEOLOGICAL ORIENTATION

The Grays Lake quadrangle in northeastern Illinois (Fig. 1) lies within the broad Valparaiso moranic system, which here is divided into several individual ridges trending nearly north-south (Fig. 2). The Tinley and Lake Border moraines lie to the east, and the Minooka,

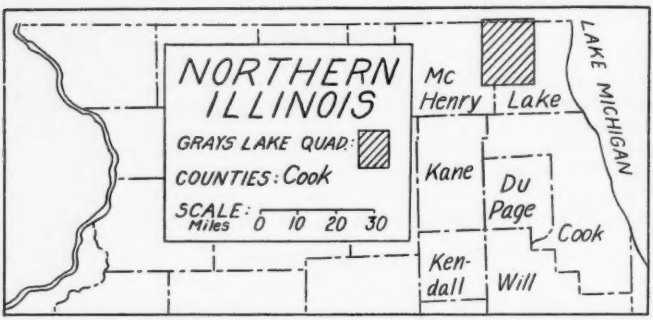

FIGURE 1.—*Map of northern Illinois*
Showing location of Grays Lake quadrangle.

Marseilles, Gilberts, and Marengo (Bloomington) moraines lie to the west. Of these the Marengo, Gilberts, and Marseilles belong to the Tazewell substage, and the others to the Cary substage of the Wisconsin glacial stage of the Pleistocene epoch.

GEOLOGIC HISTORY

The chief episodes in the late glacial history of northeastern Illinois, as previously described by Leighton (1925; 1931; 1933) and others (Thwaites, 1935, p. 82-83), are as follows:

Episode 1. Successive deposition of the Tazewell (formerly "Early Wisconsin") moraines—Marengo, Gilberts, and Marseilles—and of an extensive outwash plain, mainly behind (east of) the Marseilles moraine and north of the 42nd parallel (Fig. 3). The Gilberts and Marseilles moraines mark readvances after slight recessions.

Episode 2. Post-Tazewell, pre-Cary erosion during which Fox River and its tributaries developed valleys apparently as deep as at present (Fig. 3 A). South of Cary (near the 42°15′ parallel) Fox River and its tributaries now occupy these same valleys, but farther north Fox River now wanders through the lowland of a former fosse, connecting several large lakes. Its former winding course in the high gravel plain west of this lowland is now shared by several small underfit streams which are deeply intrenched.

Episode 3. Readvance of the glacier at the Cary substage, building in succession the Minooka, Valparaiso, Tinley, and the several Lake Border moraines. In several places these moraines descend into the Fox River valley that was eroded during the post-Tazewell interval (Fig. 4). Local postglacial erosion has been much less than that during the post-Tazewell interval.

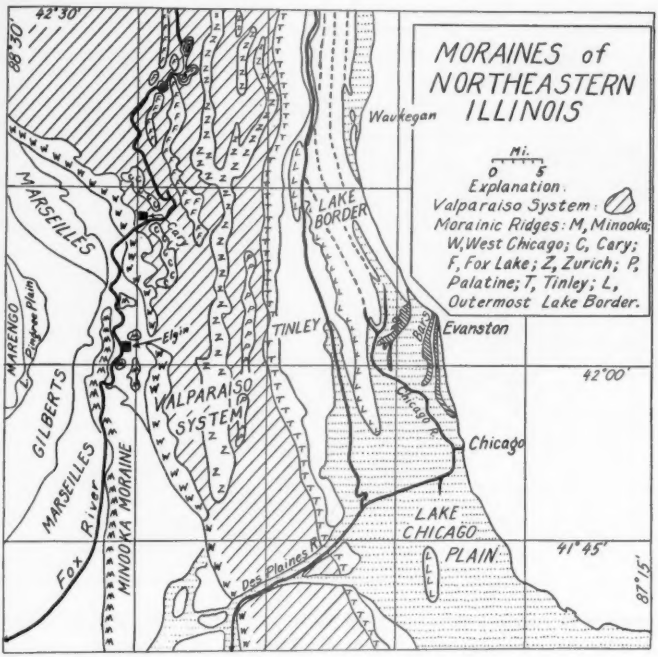

FIGURE 2.—*Moraines in northeastern Illinois*

LOCAL DEPOSITS

Subsequent studies of the glacial deposits in the Grays Lake quadrangle confirm the above history and contribute important additional information. At the northwest (Fig. 5) the Late Marseilles outwash plain constitutes an upland mantled at the west by thin Valparaiso till and farther east by outwash gravel from the narrow Fox Lake moraine of Valparaiso age, rich in kames and built along the eastern margin of the upland. East of the upland is a lowland area marking a former fosse. It is now partly occupied by several large lakes, Fox River, and associated marshes, the remainder being occupied by Valparaiso ground moraine and low terraces and lacustrine deposits of sand and clay. Fox Lake moraine and

its outwash aprons descend into this lowland farther south. Discontinuous ridges in the central and eastern parts of the quadrangle, formerly mapped as "members" of the Valparaiso moraine, are of uncertain significance, but, because outwash aprons, marginal drainage lines, and

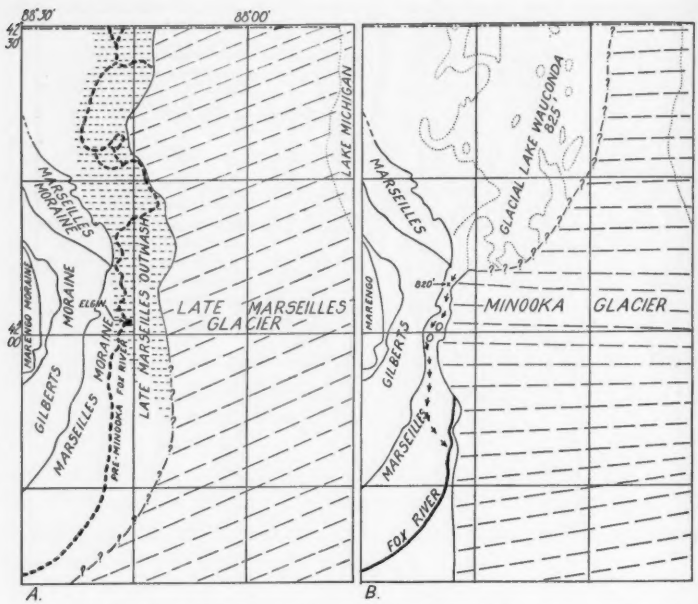

FIGURE 3.—*Conditions in northeastern Illinois during Late Marseilles and Minooka times*

A. Late Marseilles time with deposition of high outwash plain. B. Minooka time, with Glacial Lake Wauconda in Grays Lake quadrangle. Area covered same as Figure 2, with omission of Lake Chicago plain.

other features characteristically associated with end moraines are lacking, they are now interpreted as pre-Valparaiso moraines, perhaps disconnected by erosion before advance of the Lake Michigan glacier to the Valparaiso moraine.

The Valparaiso till sheet in the Grays Lake area is generally thin, indicating that it is ground moraine rather than massive terminal moraine. At numerous places the till, a sticky gray boulder clay, nowhere thicker than 25 to 30 feet, rests on older gravel and sand. Where till alone occurs, it may seem to be thicker, but, as it probably rests on older Marseilles or Minooka till with which it is nearly identical in appearance,

this greater thickness may be only apparent. The gravelly, hilly Fox Lake moraine and outwash plain at the west indicate deposition in connection with large amounts of meltwater and therefore predicate rapid melting of the glacier.

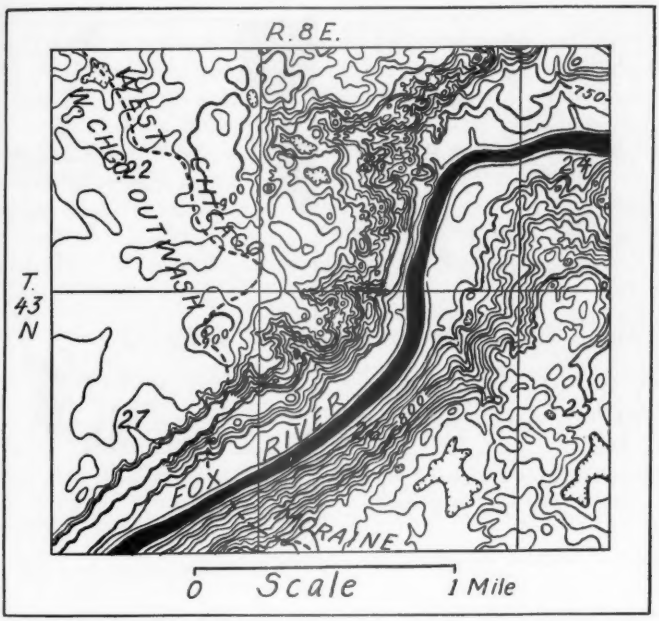

FIGURE 4.—*Topographic map of area near Algonquin, Illinois*
West Chicago moraine is shown descending into Fox River valley nearly to stream level.

A notable feature of the Grays Lake area is the presence of widely distributed deposits of very fine, unctuous, pebble-free clay, together with finely laminated silt, clay, and fine sand. These deposits are clearly lacustrine in origin and belong to two classes: (1) post-Valparaiso deposits mantling low, flat areas of Valparaiso ground moraine; and (2) deposits marked by a morainal surface topography with erratic boulders and occasional patches of till resting on the clay and silt. All deposits of (2) lie below the 830-foot level, although many are adjacent to areas that rise above 830 feet (Fig. 6). Thus they indicate a pre-Valparaiso lake at this level.

GLACIAL LAKE WAUCONDA

Both the cause and the date of this lake are apparent from a study of the Minooka glacial moraine. After the post-Tazewell erosion interval during which the Fox River valley system was cut to a depth far below

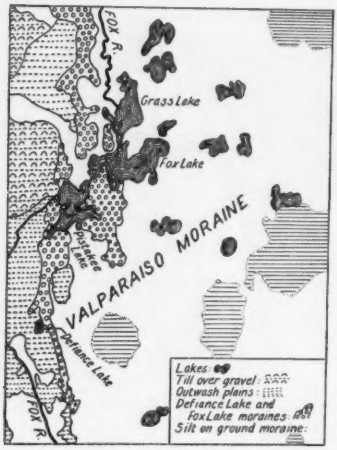

FIGURE 5.—*Significant glacial features in the Grays Lake quadrangle*

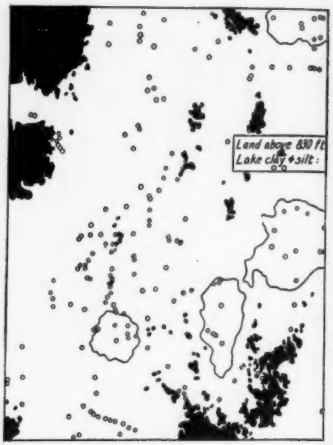

FIGURE 6.—*Distribution in Grays Lake quadrangle of water-laid silt subsequently glaciated during Valparaiso substage*

Areas above 830 feet contain no silt.

the 830-foot level, the Lake Michigan glacier readvanced into northeastern Illinois and built the Minooka moraine. In the vicinity of Elgin the glacier crossed the Fox River valley, forcing the river into a channel whose bottom now lies at 820 feet (Fig. 3 B). North of Elgin the Minooka moraine apparently trends northeast, buried beneath Valparaiso drift. A careful study of present topography indicates that this buried Minooka moraine lies near the eastern margin of the Grays Lake quadrangle. Therefore, all the region west of the Minooka moraine and below an altitude of 820 to 830 feet must have been flooded with meltwater to form a broad lake discharging through the 820-foot channel near Elgin. Wave action against islands and shores was responsible for widely scattered though thin deposits of lacustrine silt and clay. The name "Glacial Lake Wauconda" from a town in the Grays Lake quadrangle is here proposed for this lake. On these clays and silts the Lake Michigan

glacier subsequently deposited till and boulders and developed the morainal topography of the Valparaiso moraine (Fig. 2).

SUMMARY

Studies in the Grays Lake quadrangle confirm the importance of the post-Tazewell, pre-Cary erosion interval and show that at its close the upper Fox River valley was occupied by an extensive shallow lake created when the Lake Michigan glacier dammed the valley near Elgin while building the Minooka moraine. Silt and clay deposits in this lake, as well as part of the Minooka moraine itself, were later glaciated and covered by the drift of the Valparaiso moraine. The full significance of this marked change in the position of the glacial margin between the advances to the Minooka and Valparaiso moraines is as yet unknown. The conclusion of Bretz (1939, p. 52-53) that the Valparaiso moraine is a relatively thin till sheet instead of a massive moraine as formerly supposed is fully supported.

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Northwestern University, Evanston, Illinois; Illinois State Geological Survey, Urbana, Illinois.
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FIGURE 1. VIEW OF ILLINOIAN GRAVEL TERRACE
On northwest side of lower segment of Glasgow Valley.

FIGURE 2. LOOKING SOUTHWEST DOWN GLASGOW VALLEY
The Illinoian terrace is at crest of the divide near house. Main valley of the Tuscarawas is in the background.

FIGURE 3. MIDDLE SEGMENT OF GLASGOW VALLEY
Looking northeast up the valley. Level of Parker strath is saddle beyond the village. Wisconsin slack-water silt is beneath church at the right.

ILLINOIAN GRAVEL TERRACES AND GLASGOW VALLEY

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NEW EVIDENCE FOR AN ILLINOIAN GLACIAL BOUNDARY IN NORTHEASTERN OHIO

BY H. ANDREW IRELAND

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ABSTRACT

The discovery of high-level Illinoian outwash gravel in the Tuscarawas River valley near Port Washington, Ohio, proves that Illinoian ice existed in the headwaters of the present valley. The deposit lies 35 miles east of the outwash from the Illinoian drift in central Ohio. The gravel is 90 feet above the Wisconsin outwash

terraces and was deposited in a bypass of the main Illinoian drainage channel so that it was protected from removal when Wisconsin outflowing water passed down the Tuscarawas Valley. A lower gravel terrace 30 feet above the Wisconsin level is considered late Illinoian. Evidence for identifying the gravel as Illinoian, conditions

FIGURE 1.—*Location map for eastern and northeastern portion of Ohio*

during that epoch, and a discussion of the related preglacial and Pleistocene physiographic history are presented. Evidence is given to show that the Atlantic-Gulf divide at Port Washington was incised in pre-Illinoian time instead of later. The existence of Illinoian ice as far south as Shreve, Ohio, the presence of the Illinoian gravel at Magnolia, Ohio, and the gravel at Port Washington, Ohio, make

it possible to connect more accurately the Illinoian drift of the Scioto lobe in central Ohio with that of northwestern Pennsylvania. The existence of Illinoian ice across northeastern Ohio disproves statements that drainage of the upper Ohio basin during Illinoian time was through the L[illegible] Erie basin.

INTRODUCTION AND ACKNOWLEDGMENTS

Work in eastern Ohio discloses an undescribed deposit of Illinoian outwash gravel in the Tuscarawas River Valley near Port Washington, 35 miles east of Illinoian outwash material at the glacial boundary. Known drift of the Illinoian Scioto lobe lies in central Ohio east of the Wisconsin Scioto lobe in a band about 2 to 15 miles wide from near Perrysville southward to Lancaster, Ohio, from where it can be traced almost continuously into southern and southwestern Ohio (Figs. 1, 9). Most of the Illinoian drift is in basins which drain eastward to the Muskingum River. The Tuscarawas River drains westward and joins the Walhonding at Coshocton to form the Muskingum River. The Tuscarawas Valley has no connection with known Illinoian drift, and the presence of the outwash gravel in the valley proves that Illinoian ice entered the headwaters of the valley and indicates that the ice border was probably not far north of the Wisconsin drift boundary (Fig. 9).

The outwash deposit is located in the southeast portion of the Newcomerstown, Ohio, Quadrangle, about 2 miles east of Newcomerstown along U. S. Highway 36 where it leaves the main valley of the Tuscarawas River and turns northeast up a small tributary valley. (*See* arrow on Figure 2.) The road crosses a low divide, which forms the crest of the Illinoian terrace, and re-enters the main valley along a small tributary about 2 miles southwest of Port Washington. Starting about one mile west of Port Washington is a valley 3 miles long trending southwest parallel to the Tuscarawas Valley. During preglacial time the valley was a single drainage basin which drained into what later became the Tuscarawas Valley (Fig. 3). The village of Glasgow is near the center of the valley which is designated as the Glasgow preglacial valley. Stream piracy divided Glasgow Valley into three drainage segments—lower, middle, and upper; these are marked a, b, and c respectively on Figure 4. The effect of piracy is very important as it provided a bypass valley as a side channel away from the main Illinoian channel in which outwash material was deposited. The material was protected in the bypass from removal by Wisconsin outflow water.

The writer wishes to express appreciation for the comments, excellent suggestions, and valued criticism given by L. G. Westgate of Ohio Wesleyan University, W. S. Cole of Ohio State University, G. W. White of the University of New Hampshire, and C. F. S. Sharpe and C. W. Thornthwaite of the Soil Conservation Service, all of whom read the manuscript.

PRE-ILLINOIAN HISTORY

GENERAL STATEMENT

Before the description and the evidence for the age of the Illinoian outwash at Port Washington are given, it is desirable to present the pre-

FIGURE 2.—*Topographic map of southeast corner of Newcomerstown, Ohio, Quadrangle, showing the vicinity of Port Washington*

Illinoian physiographic history of the surrounding region and to mention especially the drainage changes of the Deep Stage period.

PREGLACIAL HISTORY

Port Washington is exactly on the preglacial Atlantic-Gulf divide (Coffey, 1930). The drainage northeast of the divide flowed northward up

the present upper Tuscarawas Valley and down the present Cuyahoga River Valley through the site of Cleveland, Ohio. The present Stillwater Creek, part of the Tuscarawas River, and the lower part of Sugar Creek valleys formed one major tributary, and the present Conotton Creek and part of the Tuscarawas Valley formed another (Fig. 1). The drainage west of the divide followed the old Newark Valley through Coshocton and Newark, Ohio (Tight, 1894; Lamborn, 1932). This old valley is now represented by the lower Tuscarawas River, upper Muskingum River, Wakatomika Creek, and upper Licking River valleys, and a portion is buried under glacial drift. It joined the old Teays Valley a few miles south of Columbus, Ohio, which drained northwest through Fort Wayne, Indiana (Ver Steeg, 1936). This drainage continued until the end of the Parker strath cycle of erosion.

The Parker strath represents an interrupted cycle of erosion which cut about 200 feet below an earlier erosion surface. Grade was reached in the major valleys and major tributaries, and lateral planation was beginning to cut back into the uplands when the cycle was interrupted. Meandering valleys and silt deposits of Parker time are quite marked, and the valley floor of the surface may be identified in many places, especially along the old Teays Valley and its preglacial tributaries.

PRE-ILLINOIAN ICE SHEET

Pre-Illinoian glacial material has not been found in Ohio, but the physiography and the drainage changes indicate that the older ice sheet entered Ohio and stopped north of the Wisconsin boundary. Leverett (1934, p. 94-100) has recognized the larger portion of old drift in northwestern Pennsylvania as Illinoian debris over an older drift (Fig. 9). The pre-Illinoian ice sheet has been recognized by several writers (Ver Steeg, 1934; White, 1934; Stout and Lamb, 1938) as having blocked the westward and northward-flowing streams in Ohio, causing the melt water to pond and overflow across the lowest divides to the south thus establishing many changes in drainage.

In the Tuscarawas Valley melt water from the pre-Illinoian ice plus the normal precipitation during the period was ponded between the ice front and the Port Washington divide. Water overflowing the divide undoubtedly lowered it and possibly cut it to grade since the Deep Stage drainage passed through it.

DEEP STAGE PERIOD

A knowledge of Deep Stage time is fundamental to the explanation of the site of the Illinoian gravel terrace at Port Washington and the drainage changes in eastern Ohio. The Deep Stage was a pre-Illinoian period of excessive erosion which occurred in the Ohio and Lake Erie basins.

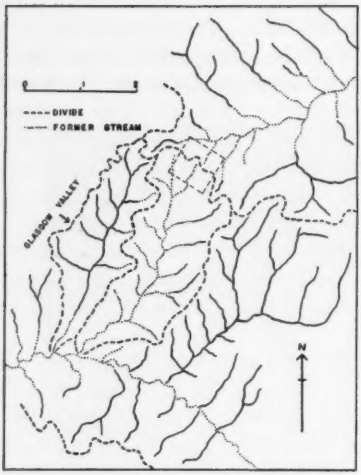

FIGURE 3.—*Probable preglacial drainage of Port Washington area*

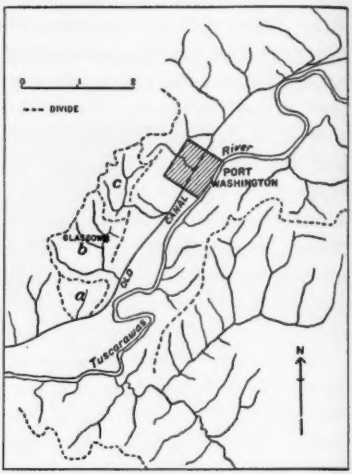

FIGURE 4.—*Present drainage of Port Washington area*

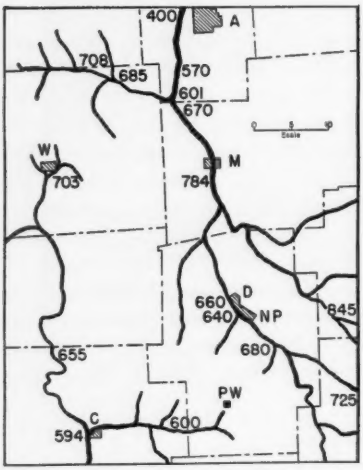

FIGURE 5.—*Deep Stage drainage channels*
(After Stout and Lamb, 1938)

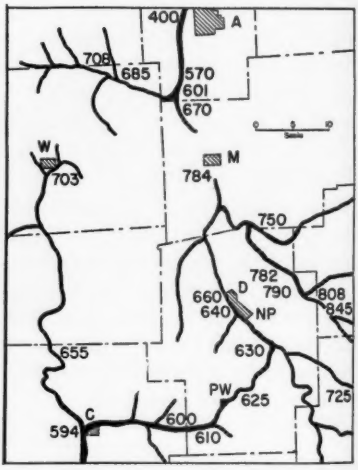

FIGURE 6.—*Revision of Deep Stage drainage channels based on additional bedrock elevations*

Downcutting exceeded sidecutting, the result being deep narrow valleys incised into the bottoms of the mature and late mature preglacial valleys. Downcutting in the major valleys amounted to 150 to 300 feet and in a few places to more than 525 feet below the present surface (Lamborn, 1932; Ver Steeg, 1934; 1936).

Several authors (Coffey, 1930; White, 1937; Stout and Lamb, 1938) place the time of Deep Stage cutting as Yarmouth interglacial. Fisk (1938) states that downcutting occurred during glacial periods when the sea level was lowered. If the time of Deep Stage cutting is taken as a transition from Yarmouth to Illinoian these opposing views are reconciled. During the climatic changes causing Illinoian glaciation it is postulated that precipitation occurred as snow in the northern latitudes and rain in the areas farther south. Between the time when the accumulation of snow began to develop a glacier and the appearance of the ice front in northern United States, there probably was a time lag during which rainfall in the more southern areas caused more active erosion because of increased gradients to a lowered sea level. The erosion probably was not so great during the retreat of the ice since the melt water was loaded with outwash debris and lacked degrading energy. If the glacial streams are regarded as aggrading, the erosion of the Tuscarawas Valley cannot be attributed to melt water.

The reduction of the Port Washington col was a prerequisite to the deposition of Illinoian outwash in Glasgow Valley. A digression is warranted to explain such cutting and the drainage changes effected, since they are related to the study of the Port Washington area and revise the present representation of the Deep Stage drainage pattern. According to the map by Stout and Lamb (1938, p. 71) the Deep Stage drainage in eastern Ohio, except Killbuck Creek, was essentially the same as the preglacial (Fig. 5). However, new evidence shows that only the Tuscarawas above Massillon returned to the northward drainage after the first glaciation, and the remainder passed through the Port Washington col (Fig. 6).

The divide at Port Washington was cut to grade level before or during the Deep Stage. This is contrary to all previous statements that the divide continued until cut by Illinoian melt water. Evidence for this is based on the elevations of the rock floor taken from the map by Stout and Lamb, well records at the Geological Survey of Ohio, core borings of the U. S. Army Engineers, and a recent well drilled at Newcomerstown.

Stout and Lamb show an elevation of 784 feet (Fig. 5) for the Deep Stage river near Massillon. Core borings of the U. S. Army Engineers for the dams of the Muskingum Watershed Conservancy District show the rock floor at Bolivar Dam to be 750 feet (Fig. 6). It is obvious that the drainage was not northward through Massillon to the Lake Erie Basin

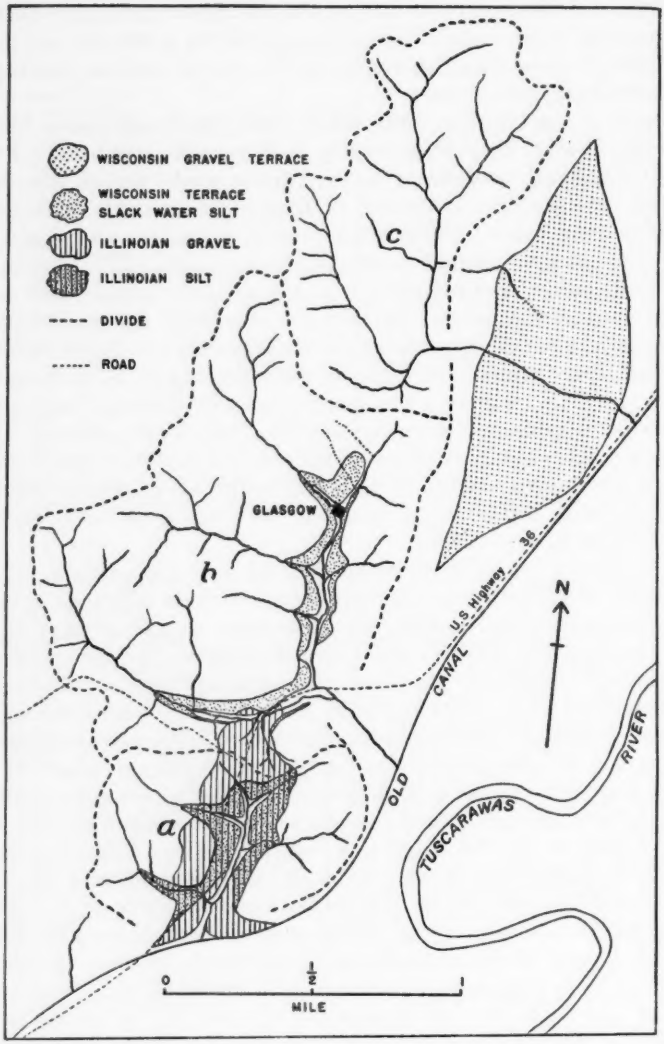

FIGURE 7.—*Detailed map of glacial deposits in Port Washington area*

across this divide which was probably created by the first glaciation. Other borings along Conotton Creek show a rise to the south with an elevation of 782 feet at the abandoned Valley Junction site, 790 feet at the abandoned New Cumberland site, 808 feet at Atwood Dam, and 845 feet at Leesville Dam (Fig. 6). This shows that the Deep Stage Conotton Creek drainage followed the northward preglacial channel as far as Navarre where it turned southward down the present Sugar Creek Valley. The elevation of 750 feet at Bolivar is consistent with the elevation of 660 and 640 feet near Dover on Stout and Lamb's map. A well 8 miles south of Dover shows bedrock at 630 feet, and one 3½ miles east of the Port Washington divide shows bedrock at 625 feet. Bedrock was encountered at an elevation of 610 feet in a recent well at Newcomerstown 6 miles west of Port Washington. This is in accord with the elevation of 600 feet shown by Stout and Lamb a few miles west of Newcomerstown. The elevation of 680 feet on their map at about the position noted above as 630 feet is inconsistent with the other very definite elevations and must be considered an error. The above elevations clearly show that the gradient of the Deep Stage Tuscarawas south of Massillon was through the Port Washington col and formed the headwaters of the Deep Stage Newark River.

The overdeepening during the Deep Stage in the Port Washington area would extend into all the tributaries as the knickpoints worked in the headwaters. There is evidence that it extended nearly to the headwaters. Borings in Sharon Valley, 10 miles northeast of Port Washington, show at least 30 feet of overdeepening within 2½ miles of the head of the valley. A well in Glasgow Valley shows 80 feet of overdeepening within 1000 feet of the headwaters of the lower segment (Fig. 8). The resistant sandstone ledges in the Hocking State Forest Parks in Southern Ohio have been incised about 150 feet below the preglacial stream bottom as a series of falls within half a mile. These conditions lead to the belief that the heads of the Deep Stage valleys were steep, narrow, or even boxhead valleys at the positions of the knickpoints.

The headward extension of the knickpoints explains the capture of the middle and upper segments of Glasgow Valley (Figs. 3, 4). The mouth of the middle segment was enlarged by the drainage from the segment. Later, the lower part of the middle segment and all of the lower segment formed a bypass from the main Illinoian outflow channel in which gravel was deposited when the Illinoian melt water was at a high level.

DESCRIPTION OF THE PORT WASHINGTON DEPOSIT

The divide between the lower and middle segments of the Glasgow Valley (Fig. 7, a and b) is composed of outwash gravel herein identified

as early Illinoian. The deposit forms a high-level terrace for 2500 feet along the northwest side and around the head of the lower segment (Pl. 1, Figs. 1, 2). The highest gravel is at an elevation of 920 feet along a side road at the crest of the divide and is at least 90 feet higher than the Wisconsin terrace of outwash gravel. The slope of the surface decreases to the southwest from 920 feet to 896 feet with an average fall of 60 feet

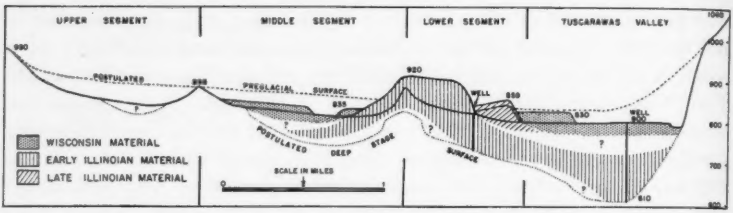

FIGURE 8.—*Projected profile of physiographic relations along Glasgow Valley and across Tuscarawas River*

per mile or less than a one per cent slope. The deposit has been well dissected by small streams to a much greater extent than observed on Wisconsin outwash terraces. The greater dissection indicates a much longer exposure to erosion and points strongly to the conclusion that the terrace is far older than the Wisconsin.

The terrace material is chiefly sand and small gravel below 2 inches in size, with occasional 3-inch cobbles. The gravel consists of well-worn quartz, granite, quartzite, gneiss, and other crystalline rock, mostly under one inch in diameter. Some limestone and hard sandstone gravel make up the larger sizes, indicating that they had not traveled so far as the more resistant crystalline rock. The Putnam Hill limestone which is exposed for 40 miles upstream is a possible source of the limestone gravel. Above the gravel lies a yellowish clay-silt generally 5 feet or more thick but absent locally.

A mechanical analysis of samples taken from a cut along the side road at the crest of the divide showed 25 per cent clay and 36 per cent silt at a depth of 2 feet. The clay decreased to 14 per cent at 4 feet and 7 per cent at 6 feet. Gravel did not occur except at 6 feet and then was 77 per cent of the sample. The higher per cent of clay in the upper 2 feet succeeded by a decrease downward may indicate a high degree of soil weathering. The sharp contact of the gravel with the soil profile at a depth of 5 feet indicates that the top material was probably stream deposited during the last stage of water in the valley. The silt over the whole area is thicker than generally occurs over Wisconsin gravel terraces and every-

where shows evidence of longer weathering. The degree of soil weathering also identifies the deposit as Illinoian. The depth of the zone of leaching is not significant as the deposit is very sandy and pervious, and any leaching has descended more than 20 feet.

The gravel surface on the northeast side of the divide between the lower and middle segments (Fig. 8) slopes rapidly to the northeast at about 425 feet per mile, and the gravel becomes finer until it grades into sand. The lower slopes are covered with Wisconsin slack-water silt at an elevation of 835 feet and above that with postglacial slope wash and slope creep from the adjacent hills. The many fragments of sandstone and shale from the adjacent hills in the deposit of Wisconsin slack-water silt indicate the source of the silty material. The eroded Illinoian surface may be traced under the younger deposits which show good contrast with the sand and fine gravel of the older materials (Fig. 8). The steep slope of the Illinoian gravel is thought to be the remnant of a valley side cut by normal slope and stream erosion during the late Illinoian and interglacial period before the advance of Wisconsin ice. The drainage from the middle segment removed the gravel which filled the mouth of the middle segment to a grade level established before Wisconsin time. The Wisconsin silt was deposited later and covered the Illinoian gravel slopes indicating the older age of the gravel. Other Wisconsin terraces of slack-water silt occur in the valley of the middle segment, rising from an elevation of 825 feet at the mouth to 857 feet northeast of the village of Glasgow, 1½ miles from the main stream.

The high-level terrace gravel was not deposited in the main channel of the outflowing Illinoian water but in a side channel or bypass formed by the capture of the Glasgow Valley during the Deep Stage period of erosion (Figs. 3, 4). The upstream portion of the bypass was along the lower valley of the diverted drainage from the middle segment, and the downstream portion was the valley of the lower segment. The small size and position of the bypass valley would admit only currents of low velocity, and the small gravel further indicates this. Wisconsin melt water did not flow through the bypass valley or it would have washed out the high-level Illinoian gravel. The present elevation of the Wisconsin outwash material shows that the water flowed at a much lower level. The high elevation of the gravel precludes deposition by Wisconsin outflow water, and such is the major evidence for identifying the high gravel as Illinoian.

Two lower terraces at the mouth of the lower segment are probably late Illinoian (Figs. 7, 8). The higher, on the southeast side, has a maximum elevation of 859 feet. The surface has a 3 per cent slope dipping upvalley, and the gravel becomes finer and gradually lenses into sand and silt within 500 feet. Along the front of the terrace the gravel averages 1

to 2 inches in diameter, but much of it is 3 inches or more, and some is as large as 8 inches. An equivalent terrace occurs on the northwest side of the segment, with the highest elevation at 845 feet. The larger sizes of

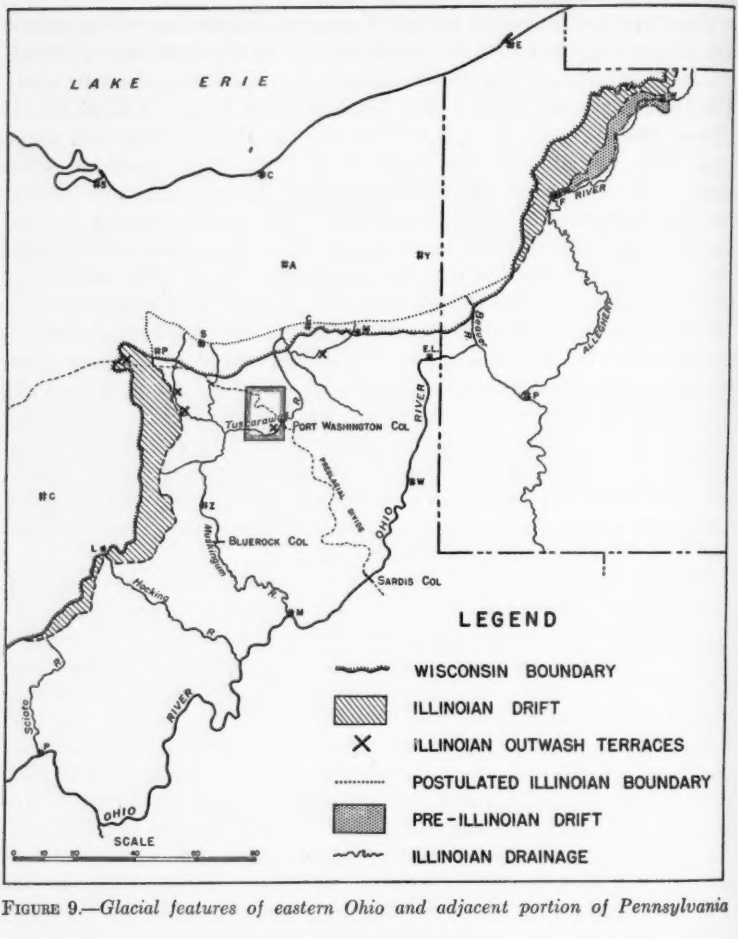

FIGURE 9.—*Glacial features of eastern Ohio and adjacent portion of Pennsylvania*

gravel indicate a stronger current than that where the high gravel was deposited. The low deposits are thought to be a valley mouth bar deposited in a back eddy from the main Illinoian outwash channel, and the large gravel grading to small away from the main channel would be ex-

pected. These terraces are too high for Wisconsin outwash and are lower than the Illinoian high gravel 2500 feet farther up the valley; they are thought to have been deposited during the lowest level of Illinoian outflow water when the divide at Blue Rock below Zanesville was reduced to nearly the present grade level. The Wisconsin terrace level in the main valley 2 miles above this point is at an elevation of 825 to 830 feet, and the slack-water terraces in Dunlap Creek Valley, 2 miles to the south, are at an elevation of 820 to 830 feet (Fig. 2). If the valley mouth bar at 859 feet is interpreted as Wisconsin, it would require the level of the outflowing water to be 30 feet higher than any of the adjacent known Wisconsin gravels or silts.

Erosion in the lower segment during the Deep Stage is shown by a well drilled on the northwest side of the valley. Bedrock was encountered at 80 feet below the surface or at an elevation of about 760 feet. The bedrock in the center of the valley might be 740 to 750 feet or lower. This well indicates that a knickpoint of the overdeepening during the Deep Stage had retreated up the valley to within 1000 feet of the divide between the lower and middle segments. This deep valley was then filled with material carried through the bypass by Illinoian water. Post-Illinoian erosion has removed much of the material—as much as 100 feet in places—down to the present level. This removal would include portions of the late Illinoian silt deposited after the bypass was abandoned.

Silt deposits occur in the upper and central part of the lower segment at elevations of 840 to 875 feet. They are too high for Wisconsin slack-water silt and they fall within the limits of the Parker strath bench. However, they might well be late Illinoian slack-water silt deposited during the lower levels of the main outwash stream. The lower segment ceased to be a bypass when the Illinoian water level fell below 920 feet, and the high-level deposit would have acted as a barrier, leaving the mouth of the lower segment as the only opening for back water from the main channel. This would account for the silt overlapping and interfingering with the low-level late Illinoian gravel deposit, as both were probably contemporaneous in time of deposition. Early Illinoian slack-water deposits would not be present, as a current then passed through the valley.

ILLINOIAN OUTWASH IN EASTERN AND CENTRAL OHIO

No remnants of Illinoian gravel, except that at Port Washington, have been found along the Tuscarawas Valley. Leverett (1902, p. 285-286) describes Illinoian gravel on Sandy Creek, a tributary to the Tuscarawas, which will be discussed below. Any high-level outwash material in the headwaters north of the Wisconsin border was destroyed when the Wis-

consin ice passed. Any Illinoian material in the Tuscarawas Valley south of the Wisconsin border was removed or reworked, except in the Port Washington area where the protection of the bypass valley prevented its removal. Late Illinoian outflow probably truncated the valley walls of the main channel, to some extent, removed all or parts of the high-level deposits along the main valley, and redeposited them at a lower level as the Blue Rock divide south of Zanesville, Ohio, was cut to grade level. Wisconsin melt water removed and reworked the late Illinoian material, any remaining deposits of the early Illinoian, and left extensive deposits now terraced by postglacial streams. Sloughing of the Illinoian material into Wisconsin slack-water areas by slope creep and erosion may have removed some of the deposition. The dearth of Illinoian outwash in the Tuscarawas Valley accords with its lack in other valleys which had connection with the known Illinoian ice front.

The Illinoian gravel described by Leverett is in a re-entrant east of Magnolia, Ohio, about 12 miles southeast of Canton, Ohio (Fig. 9). His description is brief, and additional facts seem warranted. The deposit occurs south of Sandy Creek in a low saddle which probably was the main preglacial channel of Sandy Creek. A hill of bedrock separates the saddle from the present Sandy Creek. The numerous terraces of Wisconsin age in the vicinity are at an elevation of 990 feet, while the top of the high gravel in the re-entrant occurs at 1020 to 1025 feet. The surface of the high gravel is much more dissected than the flat-topped Wisconsin terraces. Rotted granite, some with exceptionally well etched surfaces, fragments of rotted limestone, and other foreign rock are present. The gravel has been fairly well cemented for 2 or 3 feet below the surface so that it holds together in large masses and stands as a vertical or overhanging wall on the north side of a gravel pit adjacent to the Pennsylvania Railroad.

That the Deep Stage channel did not pass through the re-entrant is shown by the bedrock exposed in a strip pit for coal, where the Deep Stage channel might have been cut. The stream pattern, land forms, and valley characteristics indicate that the Deep Stage channel was cut to the north of the hill of bedrock where Sandy Creek now flows. After the deep channel was filled with Illinoian outwash, the material was spread through the re-entrant to the south which served as a bypass. Post-Illinoian erosion and Wisconsin melt water have removed the Illinoian gravel from the main channel but have left the high gravel where it was protected.

Wisconsin water may have passed through the bypass south of the main channel. This is indicated by the sharp contact, at about 990 feet, of sandy soil containing abundant small gravel with higher material of

silt and fine sandy soil containing only a few large cobbles from 3 to 5 inches in size. The higher material gives a general appearance of being older than the lower. Abundant small gravel, averaging less than one inch in diameter, along the lower slopes of the area indicate that a relatively gentle current deposited them. It is believed that this current excavated the lower part of the Illinoian material but left most of it on the north edge of the re-entrant adjacent to the bedrock core of circumdenudation.

The Illinoian terraces in central Ohio, near Bellville on the Clear Fork of the Mohican River in Richland County, near Cavallo on the Mohican in Coshocton County, and one at Walhonding near the mouth of the Mohican are high above the Wisconsin terraces, in places 100 feet higher (Figs. 1, 5). They decrease in height downstream from 1200 to 1220 feet at Bellville to 940 to 960 feet at Walhonding (White, 1937, p. 17-18). The terrace at Walhonding is 17 miles west of the juncture of the Walhonding Valley with the Tuscarawas Valley at Coshocton. The terrace near Port Washington is at an elevation of 920 feet 17½ miles east of the juncture. Slope profiles drawn from these two points to the same elevation at the juncture of the two valleys would have almost identical gradients. The position and elevation of the Port Washington high terrace described as Illinoian thus accord with a terrace previously established as Illinoian and corroborate the identification of the Port Washington terrace as Illinoian.

ILLINOIAN SLACK-WATER SILT

The lack of Illinoian slack-water silt in tributaries to the Tuscarawas basin in contrast to the widespread Wisconsin silt requires explanation. If deposited it should have remained in the tributary valleys high above the Wisconsin level since outflowing melt water could not have removed it. The filling of the enormous space left by the removal during the Deep Stage would require an equivalent amount of material derived either from the hills in each basin or from glacial outwash. The debris from the melting Illinoian ice choked the valleys carrying the melt water, but the glacial debris did not accumulate in the tributary stream valleys, except in back eddies for a short distance above the mouths where the gravel surface sloped away from the main channel into the tributaries. The gravel would become finer upvalley and interfinger with the silt during deposition. This condition is exhibited on a small scale in the mouth of the lower segment of Glasgow Valley, and many examples of the same conditions exist in Wisconsin material. Except in protected areas both gravel and silt were removed during post-Illinoian erosion.

A sufficient period of time did not elapse during the Illinoian—even though it was a long period—for normal erosion to furnish enough sediment to fill the Deep Stage valleys higher than the Wisconsin terrace levels; neither was there sufficient time during the Sangamon-Peorian interglacial, nor during the Wisconsin, for each separately to have filled the overdeepened valleys. However, the accumulation of erosional debris during the combined time of the Illinoian, Sangamon-Peorian, and Wisconsin was sufficient to raise the top of the valley fill to the present level.

The Illinoian silt in Glasgow Valley is recognized only because of its association with the Illinoian outwash gravel. Some geologists might contend that Illinoian slack-water silt was deposited at a high level in other tributary valleys; but there is no evidence of this, and such silt would be difficult to differentiate from preglacial silt of the Parker strath or from post-Wisconsin slope wash as it was all derived from the same source. It seems more logical to think that it never accumulated above the Wisconsin level, except at the mouths of the tributaries, as illustrated by the material at the mouth of the lower segment of Glasgow Valley.

PLEISTOCENE EROSION IN THE TUSCARAWAS VALLEY

The Tuscarawas Valley at the Port Washington col does not narrow as it was widened by Deep Stage and post-Illinoian erosion (Fig. 2). The sharply truncated spurs, the greatly increased width, and straightness of the valley without meander bends are in great contrast to the marked meandering and interlocking spurs of the Stillwater and Wills Creek valleys which inherited their courses from preglacial drainage. Stillwater and Wills Creek did not carry glacial outflow, and the erosive power there was from local precipitation in small basins and less than in the Tuscarawas Valley where rainfall and melt water were combined.

The straightened valley is distinctly an erosion feature, though modified by filling, but the straight character cannot be attributed to the burial of landforms by aggradation. The preglacial interlocking spurs and the meander bends of the Tuscarawas Valley were developed on a grade level much higher than the level of the present Wisconsin terraces and have not been buried. The apparent increased width of the valley may be attributed to aggradation, but the actual width is due to erosion.

The bedrock in the Muskingum River at the Blue Rock col is at an elevation of 670 to 680 feet and at least 60 feet higher than bedrock at 610 feet in the Tuscarawas channel at Newcomerstown. Blue Rock col was higher during Deep Stage time, as it was lowered to the present elevation by post-Illinoian erosion. The Muskingum was the trunk drainage for Illinoian water in eastern Ohio but it does not exhibit the great

erosion on the valley sides and entrenchment during the Deep Stage as does the Tuscarawas Valley. The major erosion which straightened the Tuscarawas Valley was not Illinoian or post-Illinoian but pre-Illinoian when the Deep Stage drainage was through the old Newark Valley; otherwise the Muskingum should show similar effects. The same conditions are applicable to the Walhonding Valley and to the old valley abandoned by the Newark drainage and now occupied by Wakatomika Creek. The Newark Valley exhibits the same physiographic features and was cut, definitely, by Deep Stage erosion.

ILLINOIAN DRAINAGE

Stout and Lamb (1938, p. 75-76) believe that the Illinoian drainage from the Tuscarawas Valley passed down the Muskingum River and then up the present Ohio Valley through the col at New Martinsville, West Virginia, and Sardis, Ohio, into the Lake Erie drainage by way of the Beaver-Grand River channel (Figs. 1, 9). In view of the new evidence for Illinoian ice in the upper Tuscarawas Valley and the probable connection of that ice with the Illinoian ice in northwestern Pennsylvania, it does not seem possible for any of the eastern Ohio drainage to have flowed out to the north. Stout and Lamb express uncertainty regarding the blocking of the Cleveland River—present upper Tuscarawas and Cuyahoga drainage—by Illinoian ice. The existence of the Illinoian outwash in the Port Washington area would eliminate any uncertainty because the outwash is proof that the Illinoian ice blocked the northward drainage. Illinoian drift described by Leverett (1931, p. 59) along the shore of Lake Erie east and west of Cleveland is additional evidence for the ice obstruction to northward-flowing streams. Outwash deposits of Illinoian age have been recognized along the Allegheny and Ohio rivers from the New York state line across Pennsylvania to East Liverpool, Ohio, with decreasing elevations downstream (Leverett, 1934, p. 110-113). This deposition would require an Illinoian stream flowing southwestward and not northward.

Since it was not northward, the drainage must have been down the Ohio River. This requires explanation because the Illinoian ice blocked the Ohio River at Ripley, about 45 miles southeast of Cincinnati. There was an enormous space available for water storage in the Deep Stage valleys, and huge inland digitate lakes probably occupied the valleys of the upper Ohio basin with an outlet near the south edge of the ice front of Cincinnati. Some authors object to an outlet there because of lack of evidence for it. The channel was used for a short time only, because the ice soon melted back leaving the Ohio Valley open. The channel was

probably not well enough defined in such a short period to withstand the long period of erosion after the channel was abandoned and still leave evidence of its former existence. The ice border may have been so crevassed or attenuated that the drainage was dispersed through the peripheral area and lacked erosive power. Though a definite channel is lacking to prove the location of the outlet, the water must have drained somewhere to the west around or through the ice front because it could not drain to the north.

When the ice retreated in the Cincinnati area and opened the Ohio drainage the ponded water cut through former divides and reduced them, thus integrating several drainage systems. The two most important cols—on the Ohio River at Sardis and at Blue Rock on the Muskingum River—were reduced to a grade not far above the present one, and the present courses of the Ohio and Muskingum rivers were established.

The Illinoian melt water from Sandy Creek Valley flowed up the present Tuscarawas Valley to Navarre, Ohio, and then went southward along Sugar Creek and the lower Tuscarawas Valley. Illinoian outwash, therefore, would not be expected in the Tuscarawas Valley from the mouth of Sandy Creek south to the preglacial divide east of Dover, as that divide was not cut through until Wisconsin time. Currents capable of carrying gravel could not have passed along that part of the valley, but it seems likely that slack-water silt was deposited in the ponded area back of the Dover col as well as in the tributaries to that portion of the valley.

The Illinoian ice front was farther north than Navarre, Ohio, because the drainage from the Sandy-Tuscarawas-Conotton Valley had to pass through Navarre in order to get into the Sugar Creek-Tuscarawas channel. If the ice had been south of Navarre the ponded water would have overflowed through the Dover col, which it did not do.

POSTULATED ILLINOIAN BOUNDARY IN NORTHEASTERN OHIO

The areas which give the basis for location of the Illinoian boundary in northeastern Ohio will be discussed from west to east. The drift boundary of the Illinoian Scioto lobe south of Perrysville trends northwest until it passes beneath the Wisconsin drift (Fig. 9). White postulated that the Illinoian ice sheet extended eastward from central Ohio at a short distance north of the Wisconsin border (White, 1934, p. 377). He states,

> "To block the mouth of the Deep Stage Holmes River . . . an obstruction was necessary. It is therefore postulated that a lobe of the Illinoian glacier east of the Scioto lobe extended as far south as Shreve. . . ."

The ice, therefore, must have extended several miles east of Shreve, which is located in southwestern Wayne County, 18 miles east of the border of the Scioto lobe (Fig. 9). The presence of the Illinoian gravel at Port Washington proves that a lobe of ice entered the headwaters of the Tuscarawas River. This portion of the lobe would logically be connected through Shreve with the Illinoian Scioto lobe near Perrysville. The area of recognized Illinoian gravel at 1020 to 1025 feet near Magnolia, Ohio, suggests that the ice boundary was not far distant up Sandy Creek. The head of Sandy Creek is within a few miles of Minerva, 11 miles northeast of Magnolia, and, since Sandy Creek contains Illinoian outwash, the glacier must have entered its headwaters. Therefore, it is thought that the Illinoian ice came to within 5 to 10 miles of the Wisconsin border at Minerva. The area of Illinoian drift in northwestern Pennsylvania, about 85 miles long and less than 10 miles wide, lies about 50 miles east of Minerva. The extension of the Illinoian boundary from Minerva to the drift in Pennsylvania would reasonably connect that drift with the boundary of a lobe across the Tuscarawas Valley. These conditions are the basis for drawing the postulated boundary of the Illinoian ice in northeastern Ohio from the Illinoian Scioto lobe in central Ohio to the Lobe in northwestern Pennsylvania as indicated on Figure 9. The ice lobe is designated as the Tuscarawas lobe and is in the same position as the Killbuck and Grand River lobes of Wisconsin age.

CONCLUSIONS

The Atlantic-Gulf divide at Port Washington was lowered by Deep Stage instead of post-Illinoian erosion. A recent well and other elevations on bedrock show a gradient through the col southward from Massillon, Ohio, proving that all drainage in the upper Tuscarawas basin did not return to the northward preglacial channels after the first glaciation. The entrenchment of the Deep Stage through the col explains the presence of a bypass valley in which Illinoian outwash has been protected.

The identification of Illinoian outwash in the Tuscarawas Valley near Port Washington, Ohio, proves that Illinoian ice entered the headwaters of the Tuscarawas Valley. Direct evidence for the age of the material is the high degree of dissection by streams, the deep weathering, overlap on the Illinoian gravel by Wisconsin slack-water silt, and its position 90 feet above the well-defined Wisconsin terrace level. The accordance in gradient, position, and elevation of the gravel with previously recognized Illinoian outwash in the confluent valley of the Walhonding River constitute additional support to the Illinoian age.

Two lower terraces, 15 and 30 feet above the Wisconsin level, are probably late Illinoian valley-mouth bars deposited during the lower levels of

Illinoian outflow when the Blue Rock col south of Zanesville was cut to grade level.

Its location in a bypass channel of the main Illinoian valley protected the deposit from post-Illinoian removal. Headward erosion of knickpoints and piracy during the Deep Stage period of erosion created the bypass. Post-Illinoian erosion and Wisconsin melt water removed other Illinoian material along the Tuscarawas Valley.

The presence of a system of deep valleys excavated during the Deep Stage period explains the lack of Illinoian slack-water silt in tributary valleys. Outwash material was lacking in the tributary valleys, except at the mouths, and the amount of silt derived from local erosion within each tributary basin was not sufficient to fill the Deep Stage valleys. The level of Wisconsin slack-water silt in the tributary valleys represents the combined accumulation in the overdeepened valleys from Illinoian through Wisconsin time.

The major part of the great erosion which straightened the Tuscarawas Valley and truncated the interlocking preglacial spurs was during the late Deep Stage period immediately preceding the advance of Illinoian ice.

Illinoian water did not drain northward through the Beaver-Grand River channel. The presence of Illinoian drift at Cleveland, the outwash at Port Washington, and the Illinoian drift in Pennsylvania with the southward gradient of Illinoian terraces along the Allegheny and Ohio rivers prove that drainage to the north was blocked by ice. Drainage must have been to the west along the Ohio River. Though the Ohio was obstructed during the early Illinoian, the channel was opened as the ice retreated. Drainage for a short time may have been through or around the ice front south of Cincinnati, Ohio.

The Illinoian drift near Cleveland proves that the ice extended that far south and east. The existence of the Illinoian ice as far south as Shreve, as indicated by White, the Illinoian gravel in the Tuscarawas Valley, and the gravel at Magnolia indicate that the border was probably very close to the Wisconsin border and that it extended as far east as Minerva at the head of Sandy Creek, a distance of 70 miles. From Minerva it is 50 miles to the Illinoian drift in northwestern Pennsylvania, and the extension from Minerva would reasonably connect the ice boundary in central Ohio with that in Pennsylvania. These conditions are the basis for drawing a postulated boundary of an Illinoian ice lobe in northeastern Ohio.

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FIGURE 1. TUNGSTEN-BEARING SHALE STRATA
Thin quartz vein (center); ocherous zone above loose block.

FIGURE 2. CHERTIFIED LIMESTONE CUT BY QUARTZ VEINLETS

SEDIMENTARY SERIES

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TUNGSTEN-BEARING MANGANESE DEPOSIT AT GOLCONDA, NEVADA

BY PAUL F. KERR

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ABSTRACT

The tungsten deposit lies just above the former high water level of Quaternary Lake Lahontan and a short distance east of Golconda, Nevada. Tungsten-bearing manganiferous and ocherous deposits underlie calcareous tufa. The tungsten ores lie blanketlike on an erosion surface which truncates tilted Triassic sediments. Beneath the blanket deposits are veins of similar mineralization which are thought to have provided the source of the overlying ores. The blanket deposits are sufficiently large and contain enough tungsten to justify economic development. The ore minerals are colloidal in origin, tungstic acid having been adsorbed in psilomelane and limonite while both were gels. Both the ore-bearing layers and the tufa are considered to be chiefly of hot spring origin rather than the result of precipitation from Lake Lahontan.

A jarosite-bearing vein containing small amounts of tungsten crops out at a slightly higher elevation where the hot springs deposit has been removed by erosion. The vein cuts steeply inclined limestone which has been considerably altered to chert and highly silicified with the production of quartzose masses. This vein is thought to be indicative of a still lower level of mineralization than the veins immediately below the blanket deposits.

It is believed mineralization started with chertification and silicification but ultimately resulted in precipitation of tungsten, iron, manganese, and tufa. The source of the tungsten is not exposed, but the existence of scheelite deposits in the vicinity suggests the possibility of underlying scheelite or perhaps wolframite-bearing veins.

INTRODUCTION

Previous studies of the tungsten-bearing manganese deposit near Golconda, Nevada, are worthy of special mention. R. A. F. Penrose, Jr. (1890), while investigating the manganese resources of the United States and Canada for the Geological Survey of Arkansas, made the first recorded geological study of the locality. In the first volume of the Journal of Geology (1893) he published a more complete description including a chemical analysis of the ore and pointing to the Pleistocene age of the deposit. Most interesting of all from the standpoint of the present discussion was the report that the ore selected for analysis contained 2.78 per cent WO_3. Since ores containing even 1 per cent WO_3 have for some time been considered workable even at considerable depth, the lack of development at Golconda during the intense search for tungsten throughout the West in 1917-1918 should be explained. It may be attributed to the marked difference between Golconda ore and common tungsten ores and a greater difficulty in treatment. The ores analyzed contained 65.66 per cent MnO, and for this reason the deposit has been most frequently referred to as a manganese prospect. In fact, during the World War several shipments of manganese ore were made, although ore in sight has never been considered sufficient to justify operations for the recovery of the low-priced manganese.

The deposit escaped further mention until Harder (1910) in an examination of manganese deposits of the Great Basin region called attention to the earlier work of Penrose. Later Palmer (1918) referred to the Golconda locality and called attention to other occurrences of tungsten-

bearing manganese ores. Pardee and Jones in 1920 published a brief description with some modification of the theory of origin proposed by Penrose.

Notwithstanding the fact that the deposit had long been known, it had remained open for location until 1935 when it was located by a group of prospectors living in Golconda. In the summer of 1935, the Nevada-Massachusetts Company purchased the claims. A short time later an opportunity was afforded the writer to make geologic studies including a geologic map and certain laboratory investigations. A topographic base map was made with a plane table by Mr. Ralph J. Holmes and Mr. Arthur F. Hagner of Columbia University. Following preliminary geological investigation, a program of research was undertaken by the Nevada-Massachusetts Company in an attempt to develop a method for the extraction of tungsten from the manganese ore. Mr. Donald Read, chemical engineer, and Mr. George Crerar, metallurgical engineer, have been largely responsible for this program which has covered several years. As a result of their studies, a pilot plant for removing tungsten from Golconda ore has recently completed a period of successful operation, and a 50-ton treatment plant at Mill City, Nevada, is nearly finished. In the meantime a program of development has been carried on under the direction of Mr. Ott F. Heizer, general manager.

Recently, through a grant from the Penrose Bequest of The Geological Society of America, the writer has obtained chemical analyses of the Golconda ore and completed a series of laboratory studies.

It should be a pardonable digression to point to the role which the name of Penrose has played in connection with this unusual locality. It was through the original investigation of R. A. F. Penrose, Jr., that the nature of the deposit was first placed on record. Again, after 50 years, description of later discoveries has been aided through the assistance of the Penrose Bequest established through his generosity.

The tungsten-bearing locality (Fig. 1) lies about 3 miles east of Golconda (U. S. Geol. Survey, Sonoma Range Quadrangle) in low hills at the foot of the Edna Mountains, just south of the opening between Osgood Mountains and the Edna Mountains cut by the Humboldt River in its course along Emigrant Canyon. The Victory Highway crosses the foothills of the range 1 mile south, and about 1 mile distant on the north the transcontinental lines of the Southern Pacific and the Western Pacific railroads occupy opposite banks of the Humboldt River. The locality is easily accessible by a dirt road which follows the historic but now abandoned roadbed of the Central Pacific from Golconda. The terrane consists of moderately rounded hills over most of which an automobile may be driven without clearing a road.

GEOLOGICAL RELATIONSHIPS

Broadly considered the geological features of principal interest at Golconda are (1) a steeply inclined sedimentary series, (2) tufa deposits distributed unconformably on the eroded surface of the sediments, (3)

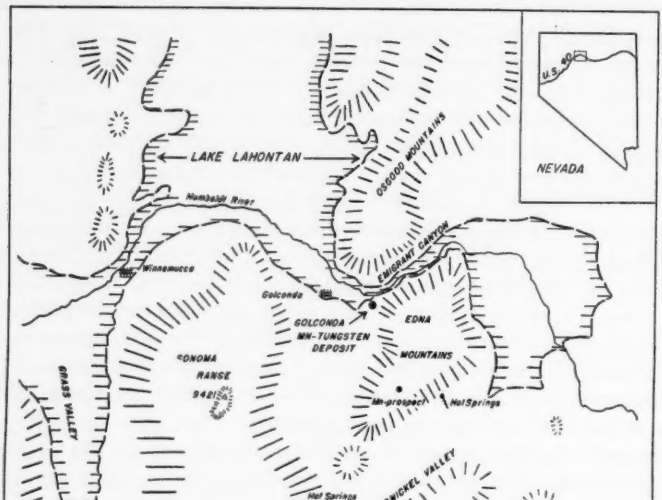

FIGURE 1.—*Sketch map of the Golconda region, Nevada*

Showing location of tungsten-bearing manganese deposit in northern Nevada. Broken lines show the approximate high water level of former Lake Lahontan.

vertical tungsten-bearing manganiferous veins in the sediments and more or less horizontal blanket deposits of similar material beneath the tufa (Fig. 2).

SEDIMENTARY SERIES

General statement.—The western slopes of the Edna and Osgood mountains are underlain by steeply inclined sediments. Locally these are capped with erosion remnants of lava, a large part of the series having been once covered with basalt. In the vicinity of the Golconda deposit, the sedimentary formations have suffered little intrusion, and there are no contact deposits. In the Osgood Mountains, however, extensive exposures of a granitic stock are to be found with tungsten deposits of the contact metamorphic type on its borders and in parts of included pendants.

The sediments at Golconda are thinly laminated and sericitized arenaceous shales (Pl. 1, fig. 1) containing occasional layers of limestone and thin beds of quartzite. Some of the limestone layers attain a thickness of 50 feet or more including intercalated shale.

The beds are more or less continuous, the general strike is northerly or northeasterly, and the inclination is steep toward the east. The dip decreases eastward from Golconda toward the crest of the Edna Mountains. Faults are numerous with displacements from a few feet to several hundred feet. Some faults may be traced for a distance of several hundred or even a thousand feet or more.

The writer made no attempt to ascertain the age of the sedimentary series. In the survey of the Fortieth Parallel, the rocks of the range to the east of the Golconda deposit were mapped as Star Peak Triassic. Other writers since have followed the same usage, although it is understood that fossils found in limestone near-by suggest that at least part of the series may be older.

Chertified and silicified zones.—Chert has formed by replacement of limestone in the vicinity of the mineralized area (Pl. 1, fig. 2). Just east of the tufa deposits, limestone beds have been substantially altered, first with the formation of chert, then with the formation of veined quartzose masses. Several stages in the replacement of the limestone are indicated both in the field mapping and in the study of specimens (Pl. 2, figs. 3, 4). Bordering the jarosite vein (Fig. 2) for a distance of several hundred feet are irregular remnants of limestone cut by masses of chert which locally effect complete replacement. To the southwest the bed is entirely chert which gradually becomes quartzose. The quartzose mass differs materially from thin spotted, slightly manganese-stained quartzite beds occurring east of the tufa area which are apparently the result of more general metamorphism.

In some places the chert is a distinct breccia, made up of unoriented, laminated, angular fragments an inch or more across. Other portions are translucent waxy chalcedony, with the typical conchoidal fracture. Chalcedony is also found in nodular masses several inches across exhibiting the pale bluish color and translucency common to the mineral. Some of the breccia has suffered further silicification resulting in lenticular masses intricately veined by quartz. The chert is occasionally stained black suggesting a connection with the manganese precipitation. This feature coupled with common hot spring precipitation of colloidal silica in Nevada suggests that silicification was an introductory phase to deposition of tungsten-bearing manganese.

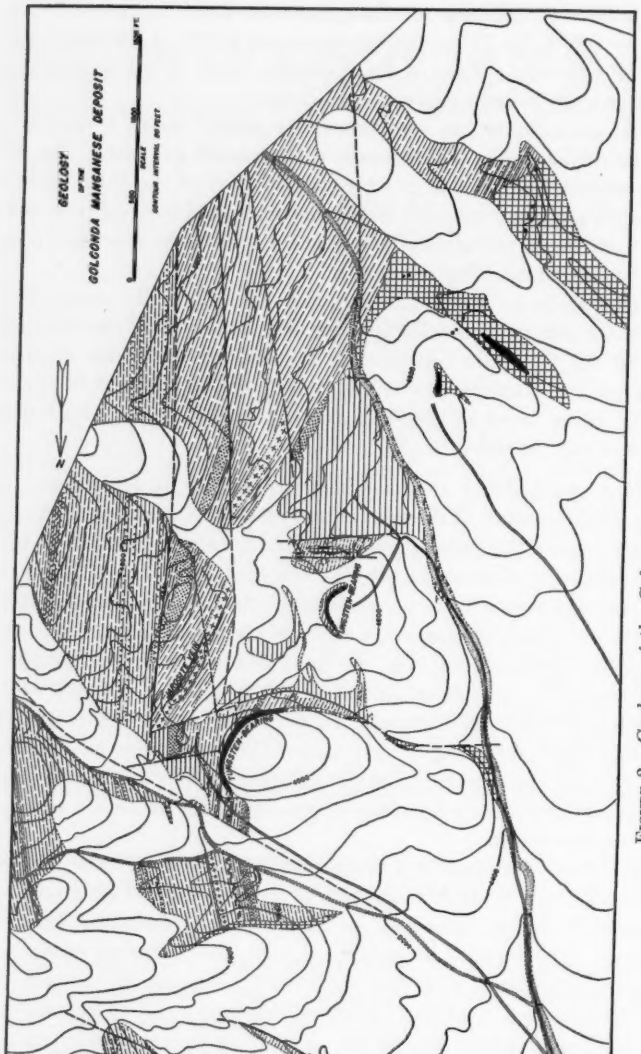

FIGURE 2.—*Geology of the Golconda manganese deposit*

Showing areal distribution of sedimentary strata, tufa, and tungsten deposits as exposed in areas where gravels have been eroded away.

METADIABASE DIKE

A few small exposures of metadiabase occur along a fault just east of the hot spring area. These probably represent a small dike cutting the sedimentary series. No other igneous rock has been observed in the immediate vicinity.

The metadiabase contains much evidence of hydrothermal alteration superimposed upon relict outlines of diabasic structure and remnant feldspars. Relict outlines of euhedral crystals, probably originally pyroxene, enclose aggregates of serpentine and chlorite. Plagioclase, probably labradorite, has been replaced largely by fine aggregates of tremolite. Mica, probably biotite, has been altered largely to chlorite with the separation of ferruginous material now largely limonite. Pyrite crystals have been oxidized to limonite. The only original minerals to be observed in thin section are a little quartz, some orthoclase, and occasional remnants of unaltered labradorite. Traces of oxidation resembling some of the oxidized material associated with the blanket deposits may be observed in thin section. Tungsten determinations made on the rock itself and on a heavy mineral concentrate by Mr. A. M. Smoot of Ledoux and Company were negative.

TUFA DEPOSITS

Form of the tufa.—Tufa deposits are well known throughout northern and northwestern Nevada, the tufa being derived either by precipitation from Quaternary lakes or from spring deposition. Russell (1885) has described the deposition of tufa in Quaternary Lake Lahontan in considerable detail. Calcareous deposits at Golconda have been thought by some to correspond to the lithoid tufa of Russell. Since the origin of this tufa has a bearing upon the interpretation of the processes involved in the formation of the manganese-tungsten mineralization, it merits special consideration.

The Golconda tufas are of two generations, an older tufa (Fig. 3), which forms partly eroded more or less fan-shaped areas varying from a few square feet to 20 acres in extent, and a younger tufa which varies from a few inches to several feet in thickness and is deposited in the depressions between the higher elevations capped by the older tufa. Both overlie the eroded upper surface of the Triassic sediments (Fig. 5). The fans of older tufa slope toward the base, where the calcareous deposition is mixed with loose, angular gravel fragments, forming breccia in which the angular rock fragments are held together with calcareous cement (Pl. 3, fig. 1). The older tufa is underlain by blankets of tungsten-bearing manganiferous and ferruginous material (Pl. 3, fig. 2).

The younger tufa forms a terracelike bench marked by a line of intermittent springs. There appears to be little accumulation of manganese beneath, and the problems connected with the accumulation of the tungsten deposits center around the older variety.

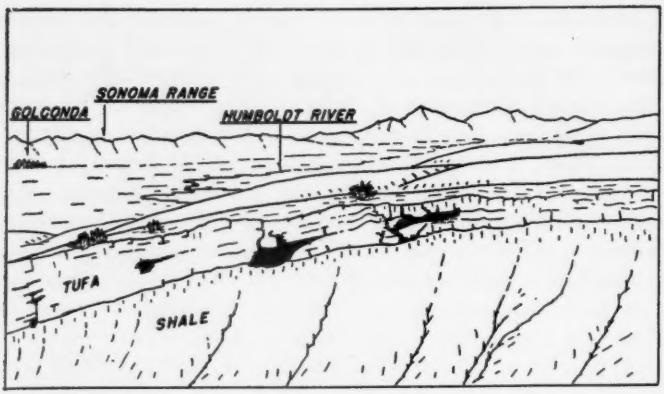

FIGURE 3.—*Sketch showing field relationships of tufa and shale*
Along eroded upper portion of a tufa cap. Valley of the Humboldt River is shown in the background.

The older tufa forms four caps or fans numbered one to four in Figure 4. Each cap slopes westerly to northwesterly from an apex, spreads out for about 1000 feet in width, and stretches from 1000 to 1500 feet in length. The deposit is thicker along the axes of the caps and at the lower margins where it may be 25 feet thick. Rimming the eroded apices of two of the older tufa caps, blankets of manganiferous and ferriferous material crop out, and a few pits penetrating the tufa have cut the underlying blankets. The caps are probably only in part underlain by manganese, especially the most northerly one which does not appear to be underlain by the manganese blanket at all. Fan No. 2 is best exposed and shows one of the best developed crescentic rims of manganiferous material. Fan No. 3 is smaller and exhibits a correspondingly smaller manganiferous crescent. Fan No. 4 is broader and less distinct but conceals tungsten-bearing manganiferous material which has been encountered in pits dug through the tufa. The known tungsten-bearing deposits lie along a belt which trends about N. 25° E. and is several hundred feet in width. Although the manganiferous blankets appear to underlie the tufa for the most part, in a few places small amounts of tufa occur beneath the manganese.

FIGURE 1. CALCAREOUS TUFA STAINED BY MANGANESE
Calcite veinlets

FIGURE 2. PHYLLITE CONTAINING MANGANIFEROUS MATERIAL IN FRACTURES

FIGURE 3. CHERT (CHALCEDONY) REPLACEMENT OF FINE CALCAREOUS SILT
X nicols

FIGURE 4. CHERT REPLACEMENT OF LIMESTONE
X nicols

PHOTOMICROGRAPHS OF SEDIMENTARY SERIES (X12)

FIGURE 1. CALCAREOUS TUFA BRECCIA
Fragments represent the sedimentary series cemented by porous calcareous tufa. (Specimen approximately 5½ by 6½ inches).

FIGURE 2. BRECCIA CEMENTED BY TUNGSTEN-BEARING MANGANESE MINERALS.
(Specimen approximately 3½ by 4½ inches).

Covering both the tufa caps and the sedimentary series are large areas of thin gravel wash largely derived from the near-by hills. Apparently most of these gravels are somewhat older than the more recent gravels of the small gullies and washes which separate the tufa caps. The debris covers such a large portion of the area that it effectively masks the outline of the lower portion of the tufa fans.

Origin of the tufa deposits.—Comments on the origin of the Golconda tufas have been offered by J. T. Pardee and E. L. Jones (1920), E. C. Harder (1910), and R. A. F. Penrose, Jr. (1893). Penrose noted the resemblance of the tufa to the lithoid tufa of Lake Lahontan as described by Russell (1885) and attributed the deposition to accumulation in one of the bays of the old Pleistocene lake. The accumulation of the manganese, on the other hand, was believed to be due to spring action.

Springs in the younger tufa are still weakly emitting water highly charged with salts. The location of the older tufa in the vicinity of the manganiferous zones is also suggestive of spring action, but it must have been much stronger to have formed such large deposits. The crude stratification and the blanketlike character of the deposit, unlike the striking tufa domes of Lake Lahontan so well illustrated by Russell and Jones, appear more representative of calcareous accumulation around springs. The lower margin of the tufa could possibly be attributed to precipitation along the margin of Lake Lahontan, but even this seems unnecessary.

Russell placed the eastern boundary of Lake Lahontan almost exactly at the position of the Golconda deposit, reporting that

> "During the Quaternary the Upper Humboldt Valley was occupied by a stream larger than the present which emptied into Lake Lahontan a few miles east of Golconda."

More recent observations indicate the substantial accuracy of this observation but suggest that the eastern limit may have extended a little farther east to the upper end of the Pumpernickel Valley through the narrow gap in the mountains formed by Emigrant Valley as shown in Figure 1.

According to J. Claude Jones (1925), the upper level of Lake Lahontan as carefully determined in a number of places was 4400 feet.[1] It was also his opinion that the level of the old uppermost shore line, as indicated by features now observable, has suffered little displacement since Lahontan time. If this is true, it hardly seems likely that more than the

[1] According to Jones (1925) the highest point of Anaho Island in Pyramid Lake is 10 feet below the highest level reached by Lake Lahontan. The low point on the divide between Mason Valley and the valley of Walker Lake was within 10 feet of the high water mark. The elevation of the high water mark is shown on a profile of terraces (p. 43).

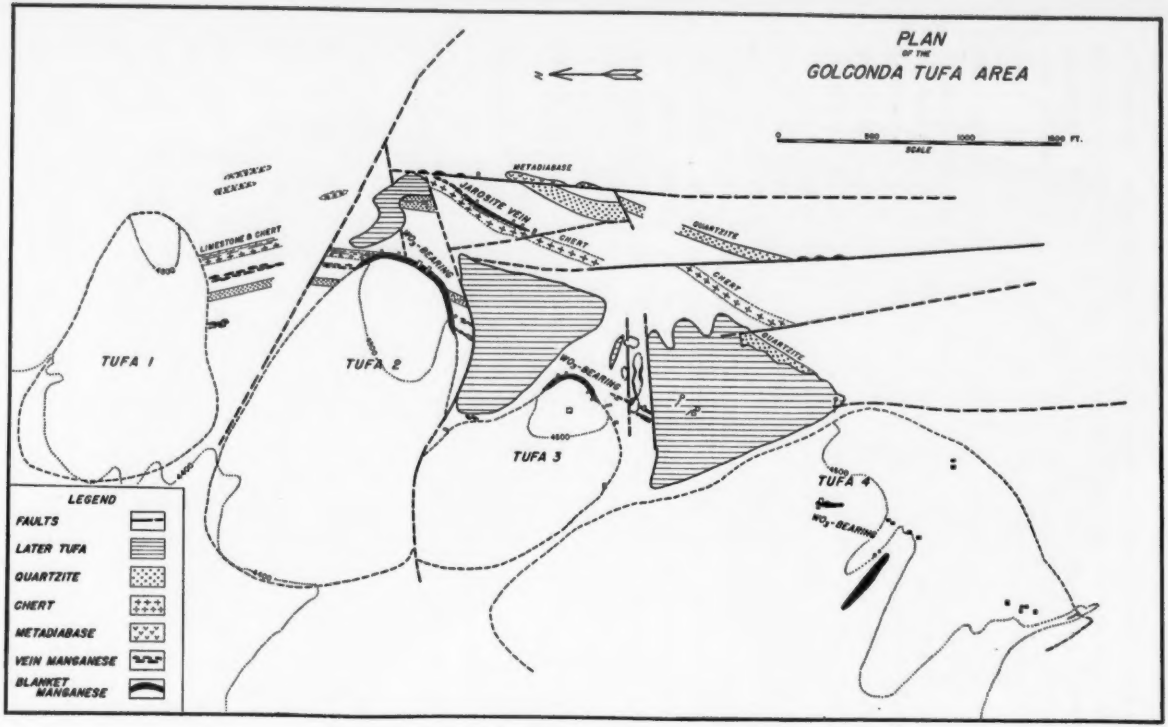

FIGURE 4.—*Plan of the Golconda tufa area showing tufa caps*

Illustrating distribution of tungsten-bearing mineralization with reference to the tufa. For convenience in description older tufa areas are divided into four units numbered 1 to 4 in sequence. Portions of the 4400- and 4500-foot contours are shown in dotted lines. The 4400-foot contour represents the highest level of Lake Lahontan as recorded by J. Claude Jones.

extreme lower portion of the tufa deposits of the Golconda locality could have been touched by Lahontan waters, since the tufa deposits as mapped extend from an elevation of approximately 4400 feet to about 4530 feet.

The topography as shown on the accompanying map (Fig. 2) has been referred for an initial datum to the near-by railroad survey elevation at Golconda with which the more recent U. S. G. S. Golconda B. M. 4386 agrees. Its greater detail confirms the more general topography of the topographic map of the U. S. G. S. Sonoma Range quadrangle. Its 4400 contour encircles the manganese tungsten locality and passes through Emigrant Valley, continuing to the upper end of the Pumpernickel Valley. Even at the highest water level as determined by J. C. Jones, the water must have been shallow in this vicinity, and allowing for the usual variations in sketching topography the 4400-foot level fell below all but the lowest portion of the Golconda tufa.

Jones has attributed the well-known erratic distribution of the tufa deposits formed in Lake Lahontan to the influence of localized conditions necessary to bring about precipitation of the tufa from the lake water.

> "The presence of the remains of algae in the dendritic and lithoid tufas, the tendency of the tufas to develop toward the light, the abundance of snails and other animal life associated with the tufas implying a food supply furnished by the algae, lead to the conclusion that the dendritic and lithoid tufas of Lake Lahontan had a similar origin. The erratic occurrence of the tufa deposits is then easily explained for they would form only where the algae became established."

Elsewhere Jones states:

> "At no time did the water of Lake Lahontan, with the possible exception of the Snake Creek Desert, become supersaturated to the point of spontaneous crystallization. As the concentration increased to the limit of solution the surplus was deposited on the tufa or solid phase with which the solution was in contact."

With this in mind, it would appear that even such portions of the Golconda tufa as might have formed beneath the waters of the lake probably owe their origin to local environment.

A number of hot springs of varying flow with or without extensive tufa deposits occur not far from the Golconda deposit at elevations both below and above the highest level of Lake Lahontan. For example, hot springs with extensive tufa occur about 7 miles to the south near the southern base of the Edna Mountains north of the Pumpernickel Valley (elevation about 4550 feet). Thirteen miles south extensive tufa deposits may be observed at Hot Springs Ranch (elevation 4621). About 16 miles west almost at the roadside of Highway 40 between Rose Creek and Lamar is a calcareous spring deposit containing a small amount of manganese and tungsten in an association similar to that at Golconda

four units numbered 1 to 4 in sequence. Portions of the 4400- and 4500-foot contours are shown in dotted lines. The 4400-foot contour represents the highest level of Lake Lahontan as recorded by J. Claude Jones.

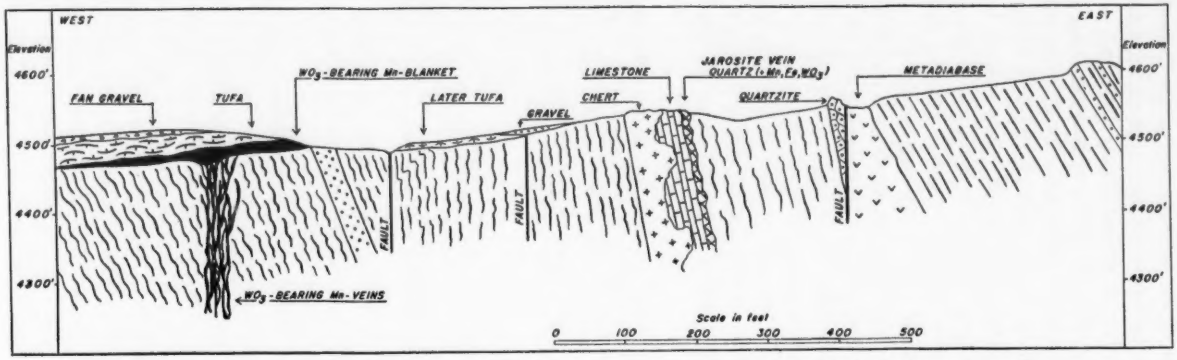

FIGURE 5.—*Cross section showing relationship of tufa and manganese deposits to sedimentary series*

Cross section is representative of area extending southeasterly from apex of tufa area No. 2 (Pl. 3, fig. 2).

(elevation about 4400). Leach Hot Springs, now yielding a moderate flow but forming only a small amount of tufa, lies 25 miles south along the eastern side of Grass Valley (elevation 4631). Just below the highest lake level at the town of Golconda are hot springs mentioned by Penrose, but deposition at this locality is small although the water flow is considerable.

Erosion has left a number of the tufa caps perched on hill crests with intermediate valleys of considerable magnitude. At lower elevation, however, the topographic irregularity above the old lake level merges into a more regular terrain at the level of Lake Lahontan. According to both Russell and Jones, the Basin Range region has suffered little erosion since Lahontan time. On the other hand, the sculpturing of the tufa caps above the old lake level at Golconda has been considerable, which suggests erosion either prior to or contemporaneous with the earlier Lahontan stages. Tufa breccia near the lower limits of the caps may have accumulated along the comparatively shallow margin of the lake. It seems likely, however, that the erosion of the upper levels occurred before the formation of the lake or during an early stage.

TUNGSTEN DEPOSITS

FORM OF THE DEPOSITS

The occurrence of tungsten is restricted to blanket deposits and veins both lying at the base of the older tufa near-by. The tungsten-bearing blanket deposit varies from a few inches to several feet in thickness and is best exposed in the crescent-shaped rims which crop out around upper margins of the tufa caps. The blankets are inclined toward the west essentially parallel to the tufa caps although some pits disclose a considerable local steepening since the underlying erosion surface is more irregular. The material is considerably intermixed in places with the lower portion of the tufa cap but does not penetrate the tufa caps in quantity for any great distance. On the surface of the eroded Triassic sediments the blanket may consist of a breccia with angular sedimentary fragments cemented by the black material. The manganiferous blanket is by no means coextensive with the tufa; in fact it is probably confined to within 500 feet of the apices of the caps. The veins in the sediments consist of irregular more or less parallel groups of branching veinlets along a N. 25° E. line near the apices of the tufa caps and a single vein containing jarosite several hundred feet east.

A cross section through the apex of one of the tufa caps with its underlying manganiferous blanket and vein system is shown in Figure 5. Only the apex of the tufa cap is shown.

JAROSITE VEIN

Several hundred feet east of tufa caps 2 and 3 at a slightly higher elevation is the vein bordering limestone and chert referred to briefly by Harder (1910) as a contact metamorphic zone. More detailed study indicates that the vein is a dense mass of jarosite, quartz, barite, calcite, limonite, and psilomelane. Assays indicate a tungsten content of about 1 per cent which is attributed chiefly to the colloidal hydrated ferric oxide carrying tungsten. Jarosite occurs as microscopic crystals which exhibit sharp rhombic outlines in thin section. These appear to be derived from the breakdown of an earlier mineralization as indicated by relict outlines of former crystal boundaries. The destruction of the earlier material has resulted in the formation of limonite which occurs in veinlets and streaks together with euhedral jarosite. Manganese has been largely replaced by iron. Enclosing the jarosite are veinlets of quartz, barite, and calcite. The barite forms minute microscopic rosettes embedded in quartz and jarosite and occurring in crystals lining small vugs. Barium is a common element at Golconda, occurring in a barite vein and as a constituent of psilomelane.

The original minerals of the jarosite vein would be expected to give a clue to the mineralization at depth. The iron sulphate content with the accompanying tungsten would suggest pyrite and wolframite, hubnerite, or scheelite. Barite is an occasional constituent of tungsten veins but is not usually precipitated under the conditions prevailing when most hubnerite or wolframite masses originate. In Arizona, eastern Nevada, and eastern Idaho, quartz, hubnerite, sericite, and traces of fluorite form perhaps the most common association in tungsten veins followed at times by metallic sulphide mineralization. If such a vein were to exist at depth, the amount of barite would be expected to decrease while hubnerite and potash mica in quartz would increase. The Golconda region, on the other hand, lies on the eastern margin of a belt in which scheelite is generally found both in veins and in contact metamorphic deposits. Near Golconda scheelite occurs in the Osgood Mountains to the north and in the Adelaide Mine to the south. Quartz is almost universally associated with scheelite, while pyrite is frequently present, and occasionally barite may be found.

The jarosite vein has been more deeply eroded than the manganese veins to the west, and any connection with blanket deposits has been removed. The tungsten, barium, potassium, manganese, and iron content of the vein, on the other hand, identify it as part of the general sequence of mineralization.

The history of the jarosite vein would appear to be as follows:

1. Thermal solutions rose along fractures converting the silty phases of the bordering limestone to chert.

2. Continued invasion of solutions resulted in quartz veins in some parts of the chert and quartz filling in a more prominent fracture bordering the limestone on the east.
3. Tungsten-bearing solutions flowed out on the surface laying down a ferruginous and manganiferous blanket containing tungsten.
4. Final stages in spring activity produced the tungsten-bearing limonite-jarosite mineralization in the vein.
5. Later solutions caused further spring action resulting in surface distribution of tufa over the blanket.
6. Erosion removed the upper portion of the vein, the bordering rocks, parts of the adjacent tufa, parts of the blanket deposition, and was accompanied by weathering.

MN-FE VEINS

Groups of small anastamosing veinlets may spread out locally through a width of 15 or more feet. These carry tungsten-bearing manganiferous material which fills the cracks and fractures in the sediments beneath the tufa caps and in eroded intermediate areas. One trench sample yielded on assay 0.33 per cent WO_3 for a width of 15 feet. Apparently the tungsten mineralization is confined to either manganiferous or ferruginous portions of this zone. One highly ferruginous streak 12 inches thick averaged 6.33 per cent WO_3. On the other hand, streaks of manganiferous material free from enclosing rock yield comparable assays. A quartz vein several inches in thickness, accompanied by branching veinlets and containing manganiferous streaks, averaged only 0.35 per cent WO_3. The shale is virtually free from tungsten mineralization except where it has been fractured and impregnated with manganiferous or ferruginous veinlets. The ferruginous material is essentially hydrated ferric oxide with accompanying quartz, some barite, and probably jarosite. The higher concentration of tungsten values in this zone may be due to greater original mineralization or perhaps to enrichment although the latter has been recorded as unusual in tungsten ores.

The veinlets which penetrate the sedimentary series contain psilomelane and abundant calcite. Quartz veins or veinlets in the same area appear to be earlier than the principal manganiferous invasion. Ferruginous masses appear to be somewhat later, apparently produced in part by replacement of the manganiferous phase. The ferruginous zones frequently contain negligible manganese, while at the same time the tungsten content may lie between 6 and 7 per cent.[2] In the fracture filling it would appear that the carbonate phase was related to the manganese phase but occurred slightly later. Nearly all manganiferous veinlets, however, contain carbonate which is easily visible in thin section but not apparent in the hand specimen. Around the margins of such veinlets, thin sections

[2] Based upon determinations by well-known assay firms and kindly confirmed by Mr. Earl Emendorfer.

indicate alteration of the shale, the muscovite of the shale being changed to a mica of much lower birefringence and probably less elasticity as indicated by warping and crenulation. Although this material is so intermixed with quartz that pure samples are virtually unobtainable, partial concentrates studied optically and by means of x-rays indicate that the material corresponds to a hydromica of the type described by Galpin (1912) and referred to by Ross and Kerr (1931). It is of the general type described as illite by Grim (1937). Presumably the thermal alteration accompanying the veinlets is responsible for this marginal change.

HISTORY OF VEIN FORMATION

Pardee and Jones (1920) outline the origin of the manganese deposit as follows:

> "A quartz lode formed along a N. 25° E. fracture along a bedding plane in the steeply tilted Triassic shale was exposed by erosion. Subsequently the outcrop of the lode and the adjacent surface of the shale were covered by tufa deposited by Lake Lahontan. Movements between the walls of the fracture brecciated the quartz lode and caused a dislocation of the tufa. Solutions rising through the reopened quartz lode spread out horizontally as soon as the surface between the shale and the tufa was reached, and by replacing the tufa, deposited the flat-lying bodies of manganese and iron oxides."

Detailed mapping indicates that the theory of origin advanced by Pardee and Jones should be somewhat modified. Although the terrain is faulted it does not appear that any one fault may be traced through the entire deposit. Fractures along the observed N. 25° E. trend provided conduits for the formation of thin quartz veins and branching networks of manganese-bearing veinlets, but the existence of a single quartz lode across the area is doubtful. The trench already described, dug along the wall of a shallow gulch crossing a network of manganese veinlets 15 feet in width, disclosed a broken quartz vein several inches thick. In another trench less than 100 feet distant, along the same strike, the network of manganese veinlets had virtually disappeared. Other trenches indicated similar variation and erratic distribution; also, where the manganese veins occur beneath the tufa cap, recent trenching has disclosed negligible displacement of the overlying cap. Careful prospecting also has failed to reveal penetration of the caps by manganese solutions other than in thin seams along small cracks or joints. Quartz veins occur along with manganese and carbonate-bearing fissures in a number of places, and quartzitic masses are associated with chertification. On the other hand, microscopic examination (Pl. 4, fig. 1) indicates that the deposition of the manganese occurred subsequent to the formation of the quartz and was followed by the deposition of calcite. Thus, although detailed studies support earlier opinions regarding a connection between the rising solutions and the blanket deposits, it would appear that the manganiferous

FIGURE 1. FRACTURED QUARTZ VEIN FILLED WITH MANGANESE.
X nicols (X 12).

FIGURE 2. MANGANESE MINERALIZATION IN PHYLLITE
(X 12).

FIGURE 3. PSILOMELANE-HOLLANDITE VEINLET
Later calcite. X nicols (X 12).

FIGURE 4. JAROSITE VEIN
Jarosite (light gray) around black limonite centers. White chiefly barite and quartz
(X 65).

PHOTOMICROGRAPHS OF MANGANESE MINERALS IN THIN SECTION

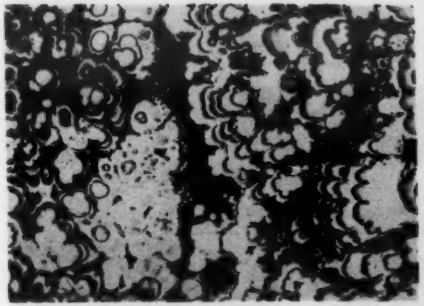

FIGURE 1. TUNGOMELANE (WHITE) IN CALCAREOUS TUFA (X 70).

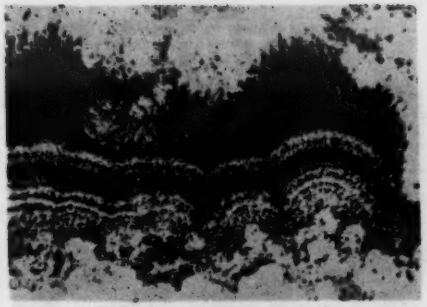

FIGURE 2. BANDED HOLLANDITE (WHITE) AND TUFA (BLACK)
Darker gray, tungomelane (X 65).

FIGURE 3. FRACTURED TUNGOMELANE (GRAY)
With streaks of hollandite (white) and tufa (black). (X 68).

FIGURE 4. HOLLANDITE AND TUNGOMELANE (WHITE) IN SHALE (X 70).

PHOTOMICROGRAPHS OF POLISHED SURFACES OF TUNGSTEN ORES

blankets were in a large measure distributed over the surface of the eroded Triassic sediments as an early phase of spring action prior to the formation of the tufa.

It is believed that the vein system in the shale is of hypogene origin. The hypogene solutions flowed out on the surface largely prior to the formation of the tufa. The temperatures at which this took place would appear to have been entirely within the hydrothermal range. Among the more significant minerals of the immediately bordering phyllite are sericite, chlorite, and hydromica, all tending toward the lower temperature end of the hydrothermal stage. Near the end of the manganese precipitation and largely prior to erosion, supergene solutions may have caused some enrichment of tungsten in the limonite zones. Probably a considerable proportion of the manganiferous and ferruginous blanket as well as overlying tufa have been removed by erosion.

With the above observations in mind, the vein mineralization at Golconda may be considered the final stage of hydrothermal activity covering the transition from fissure filling to hot spring deposition. The quartz veins, chertification, jarosite vein, and manganese-tungsten deposition would appear to be related, all occurring within a belt approximately 1 mile in length and about a quarter of a mile in width.

BLANKET DEPOSITS

A diagram illustrating the side of a prospect pit shows the character of the blanket deposition of both the ferruginous and manganiferous tungsten ore (Fig. 6) in a total section of approximately 6 feet. The upper portion of the section consists of about 12 inches of highly manganiferous black material containing occasional streaks of limonite. This zone contains 4.05 per cent WO_3. Immediately below is a 6-inch lens of soft and pulverulent yellowish-brown limonitic material. It carries 3.75 per cent WO_3. Underlying this high-grade zone is a small amount of tufa, a small sand lens, and a 12-inch layer of waxy clay almost bentonitic in physical appearance. Thin seams of manganese and ferruginous material have penetrated along fractures in the waxy clay which carries 1.01 per cent WO_3. Underlying it is an ordinary grayish clay containing numerous sooty black streaks. This layer is about 15 inches thick and carries 1.46 per cent WO_3. Beneath it is weathered shale containing a considerable amount of ferruginous material. This was sampled to the bottom of the exposure in an 8-inch thickness, the sample containing 1.82 per cent WO_3. The average of the entire 65 inches sampled amounted to 2.75 per cent WO_3. Overlying the tungsten-bearing material is a weathered overburden containing tufa varying from 3 to 6 feet in thickness.

The waxy clay in the exposure contains fine quartz and in addition massive hydromica as indicated optically and by means of x-rays. The luster, it is suggested, may be due to the extremely fine state of division of the clay mineral together with its hydrous character, a feature similar

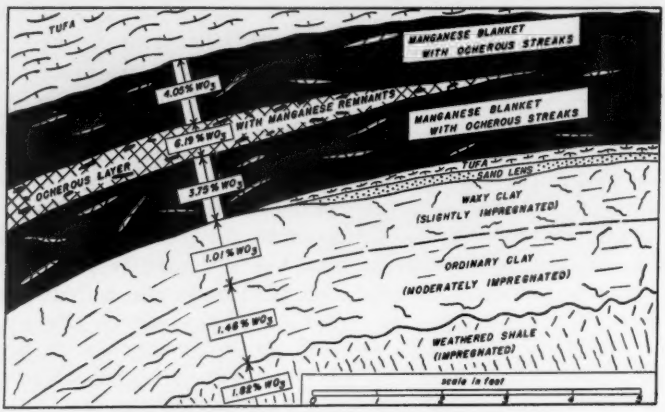

FIGURE 6.—*Diagram showing distribution of tungsten-bearing material in a prospect pit*

Dug beneath the tufa in northernmost prospects of tufa area No. 4 (Pl. 3, fig. 2).

in many respects to the waxy character of many montmorillonite layers in high-grade bentonites. Apparently the impregnation with tungsten is greater where the clay contains more quartz and is perhaps more subject to fracturing due to greater possibility of developing coarse pore spaces. The impregnation of the weathered shale is not to be confused with the hypogene veining which is present in places in which the veins are clearly marked and are much coarser than the fine streaks of the weathered shale.

The distribution of tungsten throughout the ferruginous and manganiferous ores is erratic, but, in view of the large number of assays which have been made of both ferruginous and manganiferous material covering the surface exposures and materials in a large number of pits, it seems reasonable to conclude that wherever the manganiferous and ferruginous material may be encountered in any considerable quantity, tungsten values worthy of consideration will be found. Also, the concentration in the blankets appears to be greater than in the vein. Siderite cut by veinlets of limonite has been found locally in the brown ore.

Breccias consisting of shale fragments cemented by manganese and manganese cemented by tufa occur near the bottom of the tufa caps. Since these are largely concentrated at the base and exhibit structures suggesting flowage, it is thought they represent the exudation of manganese mineralization for the most part prior to the tufa deposition. Although small veinlets of the manganiferous material extend upward into the tufa and indicate some replacement of calcareous tufa by manganese-bearing solutions along fissures, this is thought to be of minor importance. Toward the close of manganese mineralization resurgence appears to have confused the sequence of precipitation.

Several of the pits which have been dug through the tufa into the underlying manganese blanket have furnished cross sections of the tungsten-bearing material. In general they disclose that the tungsten is distributed not only within the highly manganiferous layers but in the underlying clay and shale as well, apparently due to streaks of manganiferous material in cracks and crevices and the migration of ocherous stain. Although the manganiferous material is concentrated in certain layers, and the ocherous material in others, both tend to occur together, ocherous layers containing streaks of black manganiferous material, and manganiferous layers containing streaks of brown ocherous material.

MINERALIZATION

TUNGSTEN-BEARING HYDRATED FERRIC OXIDE

Under the microscope ocherous material forms irregular fragments varying from yellowish brown to almost reddish brown, being either isotropic or birefringent with interference colors masked by the natural color of the mineral. No traces of the ordinary tungsten minerals hubnerite, ferberite, wolframite, or scheelite were observed even under the highest magnifications. Although a chemical process has been developed for extracting tungsten from the brown ore with reasonable facility chemical investigation connected with the problem of tungsten recovery has failed to contribute greatly to a knowledge of the nature and origin.

The tungsten-bearing constituent fails to correspond to any hydrated tungsten minerals on record, and x-ray diffraction patterns agree within the limits of experimental error with limonite. The material is similar to that reported by Singewald and Milton (1929) as tungsten-bearing limonite found on wolframite near Fredericktown, Missouri. Material of a similar nature has been observed on specimens of wolframite from a number of localities represented in the Columbia University collection.

The chief constituents of the Golconda brown ore identified in a sample submitted for chemical analysis were quartz, sericite, calcite, dolomite,

feldspar, and tungsten-bearing hydrated ferric oxide. Manganese is very low. The chemical analysis by Mr. F. A. Gonyer is given in Table 1.

Specimens of wolframite or scheelite may be altered to an ocherous colloidal product containing both tungsten and iron. Such a product has been observed on specimens of scheelite or wolframite from Lane's Mine,

TABLE 1.—*Chemical analysis of tungsten-bearing limonitic ore, Golconda, Nevada*

SiO_2	38.54
TiO_2	.51
Al_2O_3	12.96
Fe_2O_3	24.71
WO_3	2.64
CaO	1.44
MgO	3.05
MnO	.09
Na_2O	.78
K_2O	2.12
P_2O_5	.83
H_2O	12.19
	99.86

Long Hill, Connecticut; Meymac, France; Lead, South Dakota; and San Casmado, Portugal, in sufficient quantity to yield samples adequate for x-ray diffraction patterns. Carefully selected material in each case gives microchemical tests for tungsten and iron. It also yields weak x-ray diffraction patterns corresponding to limonite. The material is isotropic or shows birefringence masked by the red color frequently observed in limonite.

W. R. Jones (1918) stated that wolframite is frequently so weathered in tropical climates such as Burma, Siam, and the Malay States that only the hydroxides of manganese and iron or tungstic acid remain. A pure manganese mineral, or limonite, or tungstite may result. The results of weathering are not so evident in other climates as indicated by the well-preserved unaltered tungsten minerals in wolframite or scheelite placers in the western United States. The hydrated specimens from the localities mentioned above are thought to be somewhat exceptional. It would appear, however, that the unusually large amount of tungsten-bearing hydrated iron oxide at Golconda could easily result from the action of warm waters containing tungsten.

It is ordinarily recognized that tungsten and iron may occur together in colloidal hydrated form. Gannett (1919) in carrying out experiments relating to the enrichment of tungsten ores reported that

"After tungsten is dissolved, it is very easily precipitated by ferric salts; but this precipitate is colloidal, difficult to filter."

Hydrated iron minerals as reported by Posnjak and Merwin (1919) occur in two polymorphic forms—goethite and lepidocrocite—and in the so-called "amorphous" or colloidal limonite. The recognized hydrated tungsten minerals, ferritungstite ($Fe_2O_3.WO_3.6H_2O$) and tungstite ($WO_3.H_2O$), are crystalline although the crystals are frequently small. Ferritungstite has definite properties as reported by Schaller (1911) and later by Larsen (1921). Descriptions of tungstite have been given by Walker (1908; 1933) and by Larsen (1921). Although one might reasonably expect to find one or both of these minerals in the Golconda material, neither has been identified.

TUNGSTEN-BEARING MANGANESE ORE

The principal manganese mineral in the Golconda deposit is tungsten-bearing psilomelane, intimately mixed with which is a fine anisotropic manganese mineral (Pl. 5) believed on the basis of x-ray diffraction patterns to be hollandite. Pyrolusite may also be present in the deposit although the principal manganese minerals indicated by the lines on the x-ray patterns are psilomelane and hollandite.

Psilomelane has long been recognized as a complex hydrous manganese mineral. It normally contains potassium, frequently contains barium, and in various places may contain sodium, lithium, silicon, iron, zinc, lead, cobalt, nickel, copper, or tungsten. Recent studies of psilomelane from Katanga by De Leenheer (1938) have emphasized the general belief that (1) psilomelane is apt to be impure, (2) usually forms a heterogeneous mixture, and (3) is colloidal in origin. The various oxides are adsorbed by psilomelane gel, giving rise to special varieties such as asbolane, lampadite, and lithiophorite for cobalt, copper, and lithium-bearing types respectively. It has not been customary to apply any special name to the tungsten-bearing variety, the term tungsten-bearing psilomelane being probably adequate for most purposes, although, should a single word be preferred, the contraction *tungomelane* is suggested. Such a combination joins the Swedish with the Greek in a way that may be etymologically doubtful but produces a single word which is mineralogically expressive. In addition to the occurrence of tungomelane at Golconda and two other near-by localities tungsten is known to occur with manganese in the form thought to be a variety of psilomelane in Uncia mine, Bolivia, South America (Lindgren, 1922); Burma (W. R. Jones, 1918); Pleasant Valley, near Baker, Oregon (Pardee, 1922); Schneeberg, Saxony (Vaux, 1937); Sodaville, Nevada (Palmer, 1918); Tonopah, Nevada (Palmer, 1918).

The near-by localities are Rose Creek Spring near the highway west of Winnemucca and a locality known as the Hill Top prospect near the

TABLE 2.—*X-ray diffraction patterns*

Line No.	A Pyrolusite Shenandoah Valley Virginia		B Hollandite Sitapur Chindwara India		C Psilomelane and hollandite Golconda Nevada		D Psilomelane (tungomelane) Golconda Nevada		E Psilomelane Baker Oregon		F Psilomelane Sodaville Nevada	
	I	d	I	d	I	d	I	d	I	d	I	d
1	.....	.....	w	7.26	vw	7.26	vw	7.26	w	7.26	vw	7.26
2	.....	.....	w	5.00	vw	5.00	.....	.....	.....	.....	.....	.....
3	.....	.....	w	3.58	vw	3.58	.....	.....	.....	.....	.....	.....
4	s	3.13	s	3.13	m	3.14	.....	.....	.....	.....	.....	.....
5	.....	.....	w	2.76	.....	.....	.....	.....	.....	.....	.....	.....
6	w	2.72	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
7	s	2.42	s	2.42	s	2.43	s	2.43	m	2.43	s	2.42
8	vw	2.22	w	2.23	.....	.....	.....	.....	.....	.....	.....	.....
9	.....	.....	s	2.18	m	2.18	.....	.....	vw	2.18	m	2.18
10	w	2.13	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
11	.....	.....	w	1.97	.....	.....	.....	.....	.....	.....	.....	.....
12	vw	1.85	s	1.85	s	1.84	.....	.....	.....	.....	vw	1.83
13	.....	.....	w	1.74	.....	.....	.....	.....	.....	.....	vw	1.74
14	.....	.....	w	1.66	.....	.....	.....	.....	w	1.66	.....	.....
15	s	1.64	w	1.63	w	1.64	.....	.....	.....	.....	w	1.64
16	m	1.57	s	1.56	s	1.56	.....	.....	.....	.....	.....	.....
17	.....	.....	m	1.51	.....	.....	.....	.....	.....	.....	.....	.....
18	m	1.440	.....	.....	.....	.....	.....	.....	.....	.....	m	1.445
19	.....	.....	m	1.43	m	1.43	m	1.43	m	1.43	.....	.....
20	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	m	1.415
21	.....	.....	s	1.37	s	1.36	.....	.....	.....	.....	.....	.....
22	s	1.315	m	1.31	m	1.30	.....	.....	.....	.....	w	1.32
23	.....	.....	w	1.24	w	1.245	.....	.....	.....	.....	.....	.....
24	vw	1.215	vw	1.22	vw	1.225	.....	.....	.....	.....	vw	1.22
25	vw	1.130	vw	1.140	vw	1.145	.....	.....	.....	.....	.....	.....
26	w	1.065	.....	1.070	w	1.070	.....	.....	.....	.....	.....	.....
27	w	1.002	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
28	.....	.....	vw	.988	vw	.988	.....	.....	.....	.....	.....	.....
29	.....	.....	vw	.930	vw	.930	.....	.....	.....	.....	.....	.....
30	vw	.864	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
31	vw	.847	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
32	.....	.....	m	.843	m	.843	.....	.....	.....	.....	.....	.....
33	.....	.....	w	.836	.....	.....	.....	.....	.....	.....	.....	.....
34	.....	.....	vw	.810	w	.810	.....	.....	.....	.....	.....	.....

Radiation—MoKα; Intensity estimated: s—strong, m—medium, w—weak, vw—very weak.
d—measurement in Angstrom units.

southern end of the Edna Mountains. Specimens from Sodaville have been collected and studied in connection with this investigation.

X-ray diffraction patterns of the Golconda manganese ore vary between two extremes in the number of diffraction lines produced. At one are three broadly spaced lines, each considerably diffused and set in a background of strong scattered radiation. This pattern is typical of a gel-mineral such as psilomelane colloform in structure and colloidal in origin.

Specimens yielding such patterns show no evidence of crystallinity on polished surfaces even at the highest magnifications. At the other extreme are a considerable number of reasonably well developed measurable lines. This type is believed to represent a mixture of the gel-mineral tungomelane with minute crystals of hollandite. The hollandite, as well as the tungomelane, is thought to be tungsten-bearing.

X-ray diffraction patterns of hollandite, closely related pyrolusite, ordinary psilomelane, and tungomelane are compared in columns A and B of Table 2. The measurements of the hollandite shown are based on diffraction patterns of two samples, one from the United States National Museum secured through the courtesy of Dr. W. F. Foshag, the other from the Museum of Natural History, secured through the courtesy of Dr. F. H. Pough. The patterns were as nearly identical as the eye could detect, excepting the fact that the specimen from the National Museum was large enough to permit a reflection powder photograph; hence it yielded a few more lines. The X-ray diffraction pattern of pyrolusite shown in column A represents measurements of pyrolusite from the Shenandoah Valley, Virginia. Pyrolusite from Hornhausen, Nassau, Germany, and Coburg, Thuringia, Germany, gave similar patterns. Examination of the measurements recorded in columns A and B indicates that many of the reflecting planes in pyrolusite and hollandite are identical in spacing and reflecting power; a few, however, are different. Lines are present in the pattern of hollandite which do not appear in the pattern of pyrolusite and vice versa.

X-ray diffraction patterns of tungsten-bearing psilomelane from Golconda, Nevada, are shown in columns C and D of Table 2. The material in the sample of column C contains 2.24 per cent WO_3 as shown in the analysis of column 1, Table 3. The other is from known tungsten-bearing material. Both exhibit colloform structure under the microscope. The two patterns represent the extremes observed in the development of x-ray diffraction lines in the case of the Golconda material. The lines of column C are almost as numerous as the lines of hollandite, while column D contains measurements for only three lines.

If the opinion of Fermor (1917) is accepted and hollandite is the crystalline equivalent of psilomelane, pattern C would represent the transition from psilomelane to hollandite while pattern D with so many lines missing would represent psilomelane. Comparison of the figures in columns B and C in contrast to the comparison of the figures in columns A and C indicates that the transition in the Golconda material is from psilomelane to hollandite rather than from psilomelane to pyrolusite, which agrees with the conclusion of Fermor.

The nontungsten-bearing psilomelane from Baker, Oregon, is represented by the measurements of column E. Although two more lines are shown than appear in column D, no definite differences in spacing could be observed by close comparison of the patterns. The partly crystallized

TABLE 3.—*Chemical analyses*

	(1) Golconda, Nevada (F. A. Gonyer)	(2) Golconda, Nevada (R. N. Brackett)	(3) Golconda, Nevada (C. Palmer)	(4) Sodaville, Nevada (F. A. Gonyer)	(5) Psilomelane Schneeberg, Saxony (G. Vaux)	(6) Psilomelane Spitzleite, Ebenstock (G. Vaux)	(7) Baker, Oregon (F. A. Gonyer)
O.....	14.23	10.31	6.95	12.02	MnO_2–66.62	–66.73	–14.36
MnO..	68.33	65.66	48.96	61.12	7.09	7.12	65.40
BaO...	4.16	5.65	4.73	14.02	17.46	17.48	.....
SiO_2...	.21	1.70	.....	.38	.52	.51	5.42
TiO_2..	.....	.....	.....	.....	nil	nil	.09
Al_2O_3..	1.32	0.34	.....	.....	0.37	0.35	2.11
Fe_2O_3..	0.47	3.32	12.00	0.84	0.15	0.20	0.87
CaO...	0.22	3.44	1.99	0.38	0.19	0.26	2.48
MgO..	0.45	1.26	.....	1.68	0.15	0.13	0.64
Na_2O..	0.53	none	.....	0.19	tr	tr	0.52
K_2O...	3.42	0.35	.....	0.11	tr	tr	0.61
Li_2O...	.....	.....	.....	.....	tr	tr	.....
CuO...	.....	.....	.....	.....	0.48	0.31	.....
NiO...	.....	.....	.....	.....	tr	tr	.....
CoO...	0.32	n.d.	.....	.....	0.90	1.00	0.36
WO_3...	2.24	2.78	1.54	4.88	0.89	0.68	.....
H_2O...	4.04	4.16	.....	4.69	4.38	4.41	.....
H_2O...	.....	.....	.....	.....	0.48	0.50	7.35
Total..	99.94	98.97	76.17	100.31	99.68	99.68	100.21

psilomelane from Sodaville, Nevada, of column F contains 4.88 per cent WO_3 as recorded in column 5, Table 3.

The specimen from Baker, Oregon, was collected by the writer during the summer of 1939 as an example of tungsten-bearing psilomelane but on analysis was found to be free from tungsten. Tungsten-bearing material from this locality, however, has been reported by Pardee (1922). The Sodaville material was collected during the summer of 1935 and is representative of the tungsten ore from this locality. Insofar as x-ray diffraction patterns are concerned, there appears to be no observable structural difference between tungomelane and tungsten-free psilomelane. This is not suprising since considerable contamination of psilomelane has been known to occur with little if any effect upon the lines as indicated by the ordinary diffraction photograph.

Patterns of psilomelane containing unusual metals are frequently characterized by a scarcity of lines. De Leenheer (1938) studied cupro-

asbolane from a number of localities and found that all were optically isotropic and gave x-ray photographs typical of a gel-mineral. The diffraction patterns were poorly developed and consisted of a small group of more or less diffused lines. X-ray diffraction patterns of psilomelane have been recorded by Ramsdell (1932) and Smitheringale (1929). Insofar as these patterns may be interpreted, it is not clear whether they represent variations between hollandite and psilomelane or mixtures of pyrolusite and psilomelane.

Several chemical analyses of tungomelane are shown in Table 3. Analysis No. 1 represents colloform psilomelane from Golconda, Nevada, analyzed by Mr. F. A. Gonyer. As indicated by the analysis, the material consists essentially of manganese, barium, potassium, and tungsten oxides with water. Analysis No. 2 represents similar material from Golconda, Nevada, analyzed by R. N. Brackett and reported by Penrose (1893). Judging from the silica, ferric oxide, calcium oxide, and magnesium oxide content, this sample did not represent as nearly pure material as sample No. 1. Analysis No. 3 is a partial analysis of the Golconda manganese ore by Palmer (1918) which is included for comparison. Probably the sample for this analysis contained a substantial amount of limonite. Analysis No. 4 represents psilomelane similar in character to the Golconda material collected by the writer at Sodaville, Nevada. The Sodaville mineral contains more barium than the Golconda mineral and less potassium. As shown in the two analyses it contains more tungsten. However, a goodly number of tungsten determinations of the Golconda material analyzed for tungsten alone fail to indicate that the Sodaville psilomelane is higher in prevailing tungsten content. It would appear that the tungsten content of the Golconda material varies from about 1 to 7 per cent in terms of tungsten trioxide. Analyses Nos. 5 and 6 by Vaux represent psilomelane from Schneeberg, Saxony, and Spitzleite, Ebenstock. It is noteworthy that these contain small but definite amounts of tungsten and cobalt. Analysis No. 7 represents nontungsten-bearing psilomelane from Baker, Oregon, which is included for purposes of comparison. Aside from silica, alumina, and lime, this material is reasonably pure. It contains a little potassium, sodium, and cobalt, but no tungsten.

Hollandite was first recognized by Fermor (1906; 1909) in connection with his extensive study of the manganese deposits of India. The mineral was described as a crystalline manganate corresponding to the hypothetical acid H_4MnO_5. Later Fermor (1917) described crystals which he considered to be tetragonal, but subsequent studies by Vaux (1937) making use of x-rays indicate that the mineral is orthorhombic with spacings related to but not in all cases the same as pyrolusite.

Manganese specimens from Romanèche were listed as romanéchite by Lacroix as early as 1900 in the mineralogical collection of the Museum of Natural History in Paris. The romanéchite of Lacroix was thought by Fermor (1917) to be the same as hollandite, although due to the lack of description when the word was first used he did not consider it had a just claim to priority. Although romanéchite was mentioned first, the description by Lacroix (1910) in which the mineral was established definitely appeared after the original description of hollandite by Fermor (1906). The coronadite of Lindgren and Hillebrand (1904) was also considered by Fermor to be closely related, with Pb and Mn as the principal metallic elements. Coronadite was thought for a time to be discredited due to the microscopic studies of Fairbanks (1923) in which the mineral was reported to be a mixture, but more recent studies by Orcel (1931) and Lindgren (1933) serve to re-establish coronadite as a distinct mineral with the composition $2MnO_2.PbO$. Orcel has pointed out, however, that coronadite appears to represent a Pb-bearing member of the group containing hollandite, romanéchite, and coronadite. On the other hand, he does not believe that any one name should be preferred to another as a group name, in the way that Fermor (1917) proposed hollandite. According to Vaux (1937) the Romanèche material gives powder photographs which agree with psilomelane from the type locality, Schneeberg, and are distinct from patterns of pyrolusite. The powder data, however, are not recorded, and conclusions regarding hollandite are not included. Powder photographs, taken by the writer, of psilomelane from Romanèche, France, in the Columbia University collection and a specimen from Schneeberg, Saxony, from the American Museum of Natural History are identical but differ slightly from hollandite.

The mean of four analyses of hollandite reported by Fermor is compared with the mean of seven analyses of romanéchite reported by Lacroix (1910) in Table 4. The most significant feature of the analyses is the inclusion of iron, barium, and potassium oxides with the hydrated manganese oxides in the two materials. This feature was emphasized by Fermor, and it was pointed out that hollandite was to be considered as the crystalline equivalent of psilomelane. Analyses indicate that the chemical complexity of the gel-mineral psilomelane is retained in the crystalline hollandite.

Analysis (1) of Table 3 from Golconda shows similar chemical complexity. Although selected crystals of hollandite comparable to the Indian hollandite could not be obtained for separate analyses, microchemical tests, kindly made by Mr. Ford Young of Columbia University, demonstrate that both the hollandite and psilomelane are tungsten-bearing. This is to be expected in view of the chemical character of hollandite and

psilomelane. Since both are complex, if unusual elements are found in one in a certain locality, the same unusual elements would be expected in the other.

A recalculation of Gonyer's analysis of Golconda hollandite is shown in the first column of Table 4, omitting the constituents not considered

TABLE 4.—*Chemical analyses* *

	Hollandite (Golconda)	Hollandite** (Fermor)	Romanéchite*** (Lacroix)
O	14.69	13.43	12.54
MnO	70.52	69.09	65.84
Fe_2O_3	0.49	7.68	0.63
BaO	4.29	8.54	16.38
K_2O	3.53	.85	.....
WO_3	2.31	.....	.....
H_2O	4.17	.41	4.61
	100.00	100.00	100.00

* Recomputed on the basis of 100%.
** Mean of four analyses; MnO_2 recomputed as MnO and O.
*** Mean of seven analyses; MnO_2 recomputed as MnO and O.

significant. The material may be observed to correspond reasonably well with both Indian hollandite and romanéchite in manganese content but it contains less BaO. The Fe_2O_3 content is lower, and the K_2O content is higher. The water content is in agreement with romanéchite. At the present state of knowledge concerning varieties of psilomelane, it does not seem feasible to do more than suggest the possible existence of both hollandite and romanéchite as chemically related, but structurally different crystallization products. On the basis of the x-ray data the Golconda mineral is correlated with hollandite.

CONDITIONS OF MINERAL DEPOSITION

The mineralization at Golconda typifies the gel products resulting when colloidal suspensions in hot spring waters bearing manganese and iron are followed by calcium-bearing solutions. Under such conditions, the manganese and iron minerals ordinarily form gel products while calcium is precipitated as finely crystalline carbonate (Hubbard, 1922). The colloidal suspensions from which gel minerals are ordinarily produced exist under low pressure in more or less open fractures near the surface, although Krusch (1913) in a paper read before the International Geological Congress and translated by Knopf has pointed out that primary colloidal ores formed by ascending solutions presumably under more deep-seated conditions may also exist. While the related vein and blanket deposits at Golconda point to ascending solutions, these same conditions indicate predominantly near-surface deposition.

Evidence of various sorts is available to account for the source of the principal elements in the solutions—both the gel-forming and otherwise. Thus, since the fractures penetrate sedimentary strata with numerous limestone layers, it is easy to account for the source of the calcium carbonate forming the tufa. The source of the hydrated ferric oxide gel may be attributed to more or less ferruginous material in the sedimentary strata. Some of this has also formed in part by the replacement of siderite occasionally present in the blanket deposits and presumably an early mineral. The source of the manganese is less certain. Red cherts in the vicinity with some secondary manganese veinlets suggest a possible source. In a manganese prospect at the north end of the Pumpernickel Valley on the side toward Battle Mountain, substantial ledges containing pink rhodonite may be observed. Thus, possible sources are visible, but it seems fair to assume, as indicated by Hewitt and Pardee (1933), that such a mineral in substantial amount points to hypogene mineralization. Both the rhodonite and Golconda deposits belong to a belt of manganese deposits consisting of six scattered localities extending from the north end of the Pumpernickel Valley to the north end of the Osgood Range, each locality representing a different erosion level. Thus, the rhodonite deposit indicating hypogene mineralization might represent a lower level of mineralization in a sequence, with the Golconda deposit representing a higher level.

The tungsten in the gel-forming colloidal suspension would seem to form the most definite connecting link with a hypogene source. The chief problem is not so much its source but its concentration. The thermal spring in Bolivia containing 0.5 per cent WO_3 reported by Lindgren (1922) furnishes an illustration of tungsten-bearing solutions finding their way to the surface. The widespread distribution of tungsten throughout the western United States in deposits having a close association with igneous activity points to a general hypogene origin.

It is suggested that the concentration of the tungsten in the Golconda deposits may be attributed to adsorption in the gel formed from the colloidal suspension of iron and manganese. Other deposits of manganese in the vicinity are apparently more deep seated with less evidence of gel-formation and contain only traces of tungsten. Although less tungsten may have been present, part of the deficiency may be attributed to the lack of abundant adsorptive gels to take up hydrated tungstic oxide.

Although the hydrated oxides of manganese are ordinarily assumed to be of supergene origin, a number of instances have been observed in which the ore-originating solutions have been thought to be more deep-seated in origin. Consideration of these is thought to lend weight to the opinion that the Golconda ore is hypogene. In connection with the

description of orientite found in the manganese ores of Cuba, Hewitt and Shannon (1921) include a significant statement:

"The intimate association of the manganese oxides and orientite with zeolites and quartz, apparently without a pronounced interruption of the process of deposition, indicates that the first group have the same mode of origin as that commonly ascribed to the second, i.e., deposition by warm hypogene waters."

A short time later Lindgren (1922) described the tungsten-bearing manganese deposit near the Uncia mine in Bolivia. The deposits surround hot springs in an area of about 2 acres containing calcareous tufa in rude horizontal beds. Associated with the tufa are masses of black manganese ore containing psilomelane carrying tungsten. The deposition was assumed by Lindgren to be associated with the hot-spring activity.

The hollandite of India was considered by Fermor (1909) to belong to a hollandite-sitaparite-braunite mineralization produced as a result of deep-seated metamorphism. This conclusion has been further supported by Dunn (1936) who affirms that the ores are of metamorphic origin, probably on the whole with little mineral migration.

Callaghan and Thomas (1939) have described a hot-spring deposit north of Delta, Utah, which has yielded commercial manganese ore but does not contain tungsten. A bed of manganese and iron oxides scarcely more than a foot thick is mostly covered by porous calcite. The ore is reported to be a porous aggregate of psilomelane and pyrolusite in which the vugs are lined with limonite. The authors suggest that the manganese may have been leached from underlying rocks by hot circulating groundwaters.

Recently Norcross (1940) in discussing the manganese mineralization in Oriente province, Cuba, has expressed the opinion that the manganese minerals have been produced by hydrothermal activity of a regional character incidental to intrusive igneous activity.

CONCLUSIONS

After an examination of the field relationships together with a laboratory study of the ores, it seems reasonable to set forth the following conclusions concerning the origin and mineralization of the Golconda tungsten deposit.

1. The tungsten-bearing deposits are of two types, (1) more or less flat-lying blanket masses which occur along the under side of the tufa, and (2) vertical fissure fillings in the underlying sedimentary series.
2. The tungsten deposition appears to be of hot-spring origin.
3. The tufa beds overlying the tungsten mineralization for the most part are also of hot-spring origin.
4. The ores consist of limonite containing adsorbed hydrated tungstic oxide, tungsten-bearing psilomelane for which the varietal name *tungomelane* is suggested, and tungsten-bearing hollandite.

5. The adsorptive action of manganese and iron gels appears to have contributed to the concentration of tungsten from the tungsten-bearing solutions.
6. Hollandite has formed along gel cracks in psilomelane as a later part of the manganese mineralization.
7. An older, more deeply eroded jarosite vein carrying a small amount of tungsten mineralization, located near-by, appears to represent a similar vein cut to a lower level.
8. Chertification of limestones in the vicinity followed by silicification appears to be connected with the earlier stages of mineralization.

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GRAVITY SECTION ACROSS THE SIERRA NEVADA

BY W. D. JOHNSTON, JR.

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ABSTRACT

In 1939 the Coast and Geodetic Survey occupied 11 pendulum gravity stations, approximately 12 miles apart, extending from Sacramento to Reno across the northern part of the Sierra Nevada. Eight stations located upon predominantly granodiorite bedrock showed negative isostatic anomalies ranging from —4 to —44 milligals, and their average is about the same as the average of the anomalies of the 62 pendulum stations in the California-Nevada block. Three stations, located upon the belt of Carboniferous and Jurassic sediments, have positive isostatic anomalies ranging from +6 to +16 milligals. The positive anomalies are near areas of gabbro and are believed to express the excess mass of gabbro intrusives extending downward from the surface to a depth of about 5000 to 10,000 feet.

ACKNOWLEDGMENTS

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INTRODUCTION

Professor Lawson (1936) called attention to the fact that, "we have only one determination of force of gravity in the Sierra Nevada." In 1938, the Geological Survey prepared a memorandum for the Coast and Geodetic Survey outlining several gravity projects that might yield geological information of value. Among them was the request for a gravity profile from Sacramento to Reno across the Sierra Nevada. The request was kindly received by the Coast and Geodetic Survey, and, in late May 1939,

TABLE 1.—*Anomalies at gravity stations in the Sierra Nevada* (abridged from *Principal facts for gravity stations in the United States*, U. S. Coast Geod. Survey, Part 5 (1940) p. 37-38). R. W. Woodworth, Chief of Party

Station		Latitude	Longitude	Elevation	Observed gravity	Theoretical gravity	Anomaly					
							Free-air	Bouguer	Isostatic			
No.	Name								Indirect*	Depth 56.9 Km.	Depth 96 Km.	Depth 113.7 Km.
		° ′	° ′	*meters*	*gals.*	*gals.*	*mgals.*	*mgals.*	*mgals.*	*mgals.*	*mgals.*	*mgals.*
1035	Sacramento, California	38 36.3	121 27.0	11.3	980.009	980.0572	−45	−46	−19	−25	−22	−22
1036	Roseville	38 43.7	121 18.3	42.8	979.988	980.0680	−67	−72	−33	−40	−36	−36
1037	Newcastle	38 52.4	121 07.9	302.2	979.950	980.0808	−38	−71	−15	−20	−18	−19
1038	Applegate	38 59.9	120 59.4	601.6	979.914	980.0918	+8	−59	+13	+12	+10	+8
1039	Colfax	39 07.9	120 56.4	655.6	979.910	980.1036	+9	−62	+23	+25	+19	+16
1040	Baxter	39 12.7	120 47.0	1180.8	979.784	980.1106	+38	−93	+15	+22	+11	+6
1041	Yuba Pass	39 19.0	120 37.3	1745.3	979.650	980.1199	+68	−124	+8	+21	+3	−4
1042	Cold Springs	39 19.0	120 27.1	1839.3	979.597	980.1199	+44	−158	−12	+6	−17	−25
1043	Truckee	39 19.6	120 12.4	1795.1	979.583	980.1208	+16	−184	−30	−12	−35	−44
1044	Mystic	39 27.1	120 00.4	1570.2	979.630	980.1318	−17	−183	−24	−8	−29	−36
1045	Reno, Nevada	39 32.4	119 48.8	1397.3	979.688	980.1397	−21	−177	−24	−14	−28	−34

* The anomalies in this column are based on an isostatic depth of 96 kilometers, but in computing them there was taken into account the "indirect" effect, or the effect of the matter included between the spheroid and the isostatic geoid, as well as the ordinary isostatic and topographic effects.

gravity determinations were made at 11 stations between Sacramento and Reno. These stations lie on U. S. highway 40, which, in general, follows the course of the Southern Pacific across the northern part of the Sierra Nevada, traversing the Sacramento, Placerville, Colfax, and Truckee quadrangles, all of which have been mapped and published as geologic folios.

GRAVITY DATA

The principal facts for the new Sierra Nevada stations are contained in Table 1.

With those occupied in 1939, there are now 62 pendulum stations in California and Nevada. Their location and the Pratt-Hayford isostatic anomaly at each station for a depth of compensation of 113.7 kilometers are shown in Figure 1. Of the 62 stations, 56 have negative anomalies averaging —29 milligals, one has a zero anomaly, and 5 have positive anomalies averaging +9 milligals.

The algebraic average of all 62 stations is —25 milligals. In terms of the Pratt-Hayford system of isostatic reduction, this preponderance of negative anomalies indicates a general deficiency in mass of the California-Nevada block. The three positive anomalies in the Sierra Nevada with their implication of an area of excess mass present a problem of some geological interest.

GEOLOGICAL INTERPRETATION

Figure 2 shows a geologic cross section compiled from the folios indicated in Figure 1 extending from Sacramento to Reno and connecting each station by a straight line. Except on the eastern end of the section, where the Miocene andesites attain considerable thickness, the superficial rocks are not shown. The stations are indicated by black dots, and the figure above each dot is the Pratt-Hayford isostatic anomaly for a depth of compensation of 113.7 kilometers at that station. In the center of the figure the free air and isostatic anomalies are plotted graphically. At the top of the figure is shown the approximate average density of the bedrock along the plane of the section. More exact density data could be obtained, but the values given are believed to be of the right order.

From inspection of the geologic, isostatic, and density sections, certain correlations can be made.

The free air anomaly bears a direct relation to the topographic relief. Its high is centered at Yuba Pass, to the west of Donner Summit, perhaps reflecting the greater mass of the longer limb of the profile across the range.

Of the five known positive anomalies in California and Nevada, three lie on the plane of the section. They correspond approximately to the

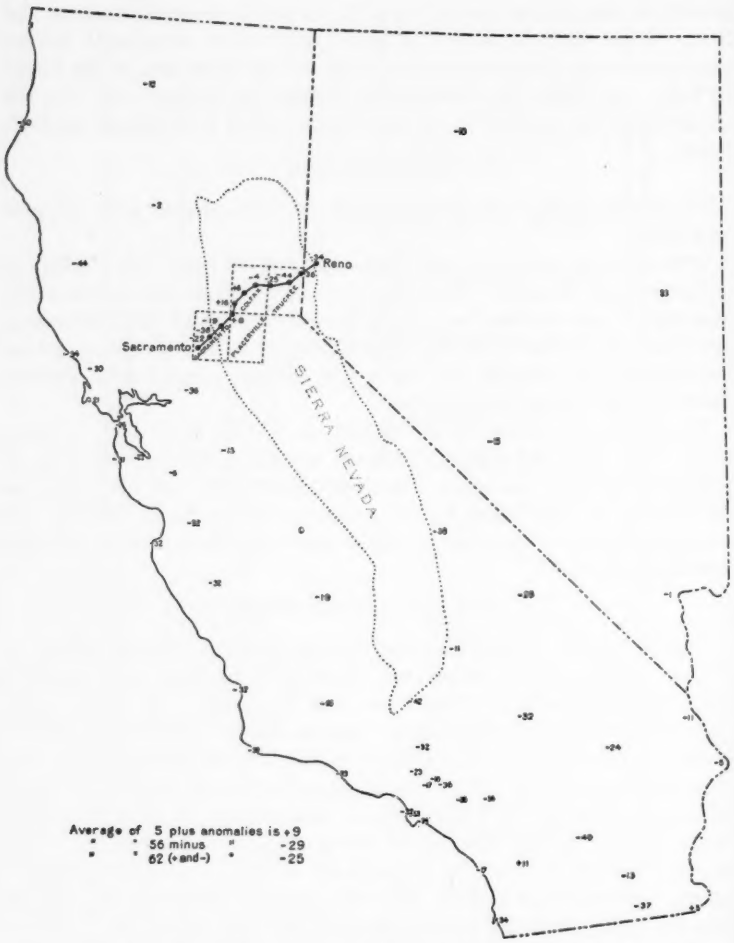

FIGURE 1.—*Location of U. S. Coast and Geodetic Survey's pendulum stations* In California and Nevada and the Pratt-Hayford isostatic anomaly at each station for a depth of compensation of 113.7 kilometers.

belt of Carboniferous and Jurassic sediments extending from Newcastle to Yuba Pass. As the sediments are lighter than the granodiorite on the flanks of the range, the positive anomalies reflect excess mass that is best attributable to the gabbro that crops out at Yuba Pass and west of Baxter,

in the plane of the section, and just to the north of the plane of the section at Colfax and between Applegate and Newcastle.

The largest positive anomaly is +16 milligals at Colfax. A cylindrical block of gabbro with a density of 2.95, having a radius of 2 miles or more

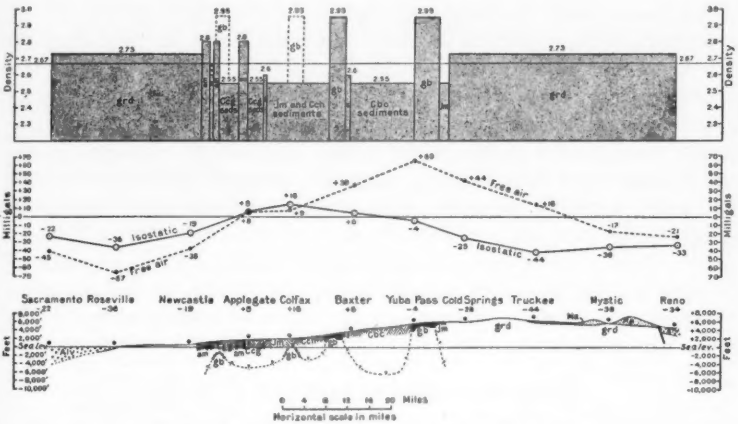

FIGURE 2.—*Geology, anomalies, and rock density*

Geologic section, isostatic and free air anomalies, and approximated average density of bedrock along plane of section.

and extending from the surface to a depth of 6000 feet, would have sufficient excess mass over the sediments which surround it to account for the isostatic anomaly of +16 milligals at Colfax. Actually, the gravity station at Colfax lies half a mile south of an area of gabbro that is 2 by 4 miles in surface extent.

Such computations, however, have little quantitative significance, for as McMillan stated in 1917:

> "From a purely mathematical point of view, any set of a finite number of observations of the intensity and direction of gravity can be satisfied not approximately but exactly by a proper distribution of the density in the earth."

These computations serve only to show that the positive isostatic anomalies along the Sacramento-Reno section are of the order that accords with assumptions regarding subsurface geology that are in general agreement with what we can actually see at the surface. The positive anomalies are near areas of gabbro and probably express the excess mass of gabbro intrusive extending downward from the surface to a depth of about 5000 to 10,000 feet.

On either side of the belt of sediments the bedrock is predominantly granodiorite, and the isostatic anomalies are negative and of about the average for the California-Nevada block.

NEED FOR ADDITIONAL DATA

It is hoped that the Coast and Geodetic Survey will furnish one or two other gravity profiles to the south across the Sierra Nevada. Further, it is hoped that some investigator in this field will attempt to refine the gravity profile between Sacramento and Reno by taking many gravimeter readings between the pendulum stations and will observe more closely the behavior of the gravity profile at lithologic contacts. If we are to learn what these isostatic anomalies mean in terms of rock mass and major structure, we must painstakingly accumulate more and more gravitational data and correlate it, step by step, with known geologic facts. Until more is known about the gravitational effect of rocks of varying density that lie on or near the earth's surface, we lack the tools to attack the broader problems of the depth of compensation or the depth of mountain roots.

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TIME AND STRATIGRAPHIC TERMINOLOGY

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ABSTRACT

In order to keep alive the effort to attain general agreement on a satisfactory dual nomenclature of time and stratigraphic units, the following classification is proposed: era—sequence; period—system; epoch—series; subepoch—subseries; chron—group; stage—formation; and substage—member. Other time and stratigraphic terms are discussed.

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INTRODUCTION

One of the most troublesome problems of scientific stratigraphic classification and nomenclature has been that of a dual classification of geological time units and the rocks formed during each. Many different schemes have been proposed during the gradual evolution of this classification, and the number of both time and stratigraphic terms in the literature is appallingly large. Numerous terms have been proposed only to fail to meet with general acceptance. Different names have been proposed at different times for the same unit; the same name has been suggested at different times for different units of either time or rock category; and the same name has been used for both time and rock unit. Authors of names have not always been consistent in the usage of the terms which they themselves have proposed. All this has resulted in confusion for those who have attempted to follow any definite, uniform scheme of classification, and most workers in stratigraphy recognize the desirability of a uniform system of classification which would have a separate time term corresponding to each recognized rock unit.

TABLE 1.—*Nomenclature adopted in 1881*

Time unit	Stratigraphic unit
Era	Group
Period	System
Epoch	Series
Age	Stage
No term	Assize

The early realization of the need for such a system of dual nomenclature is clearly shown by a brief résumé of the action of the International Geological Congress over a period of years. At the first meeting of the Congress at Paris in 1878, a committee was appointed to formulate a plan of procedure for unifying geologic classification. At the Bologna meeting in 1881 the committee recommended and the Congress adopted the scheme shown in Table 1.

Modifications of this classification were discussed at later meetings until the Eighth Congress which met in Paris in 1900 adopted the scheme shown in Table 2.

Several of these terms have persisted in common usage, but others have been more or less generally abandoned or employed with different

meaning, and new ones have come into more or less general use. Because of this condition, a scheme of dual classification which has the virtues of simplicity and familiar nomenclature is offered for consideration. None of the names proposed is new, although some have been applied with different meanings from those used in this paper. Insofar as has been

TABLE 2.—*Nomenclature adopted in 1900*

Time unit	Stratigraphic unit
Era	No term
Period	System
Epoch	Series
Age	Stage
Phase	Zone

practical, names have been retained in the sense in which they were originally proposed, except where this clearly conflicts with nomenclature established by long and general usage.

Except for the major time units, or eras, which are designated by names indicative of the general life development, most geologic time units and their rock equivalents—co-ordinates—are named from some locality or geographic feature near which they were first studied. Some, as Cretaceous and Triassic, have been named from lithologic character or subdivision of the rocks, a practice not commonly followed in present nomenclature, and a few names, like Tertiary, have been retained from otherwise obsolete classifications of geologic time and rock units.

Geologic history has been divided into major time units which have been subdivided in descending order into smaller and smaller units that are recorded by corresponding rock or stratigraphic units, each ideally characterized by a distinctive assemblage of fossils or a distinct lithology and delimited by more or less pronounced stratigraphic breaks. Different names are used for the rock units corresponding to the various time divisions, but the same geologic or geographic name is applied to both. Because of the great variation in terminology it is the purpose of this paper to propose for consideration the scheme of co-ordinate time and stratigraphic terms given in Table 3. The classification used for illustration is followed by the Illinois State Geological Survey and may not meet with the approval of everyone, especially those who consider Galena a group and its subdivisions formations. That is beside the point since it is not offered as the universally accepted classification of that portion of the Ordovician system but solely to show the proposed nomenclature.

PROPOSED DUAL CLASSIFICATION

The co-ordinate terms—phase and zone—are not included in Table 3 because a zone is not necessarily a subdivision of a member. In addition to the terms given there are several other more or less widely used names, both time and stratigraphic, which should be explained, terms

TABLE 3.—*Proposed dual classification*

Time unit	Stratigraphic unit	Geologic name
Era	Sequence	Paleozoic
Period	System	Ordovician
Epoch	Series	Champlainian
Subepoch	Subseries	Mohawkian
Chron	Group	Trenton
Stage	Formation	Galena
Substage	Member	Prosser

which because of their application need no co-ordinates. For a more complete bibliography treating of the principles of classification and nomenclature the reader is referred to *Classification and nomenclature of rock units,* by G. H. Ashley, *et al.* (1933a).

DUAL TERMS

GENERAL STATEMENT

Difficulty is often encountered in attempting to define either of two co-ordinate terms without becoming involved with the other. Insofar as possible, the following definitions or explanations have been made in terms of the time units and the definitions of the rock units fitted to the time terms. It could be argued with considerable merit that since the rock and not the time element appears to have been the dominant factor with most geologists, the order should be reversed. However, the two ideas of rock unit and time unit are so closely interwoven that a discussion of one almost automatically resolves itself into a discussion of both.

ERA—SEQUENCE

The era is the longest division of geologic time which is commonly considered. Unlike many other time or stratigraphic terms it has been used with the same meaning by practically all workers. Each era has been named to indicate the general character of the life that existed during that time. The rocks of each era are separated from contiguous strata by great breaks in the record caused by the revolutions that serve as a

basis for dividing geologic time into the more or less definite great intervals. These profound physical changes were accompanied by great biologic changes. The limits of an era should be world-wide in application.

The term sequence was proposed by Moore (1933, p. 54) for the rocks of a geologic era. Group was originally chosen for this stratigraphic unit by the International Geological Congress in 1881 and later abandoned. In more recent years it has gained almost universal acceptance for a much smaller subdivision of rocks as explained later. The word sequence is frequently employed in a much more general sense. It is customary for some writers to use it for any succession of rock strata irrespective of the magnitude or of the time interval with which the rocks may be co-ordinated. This is illustrated by the following quotations from recent publications:

"The subcommittee believes that the paleontologic and stratigraphic evidence at hand is sufficient to justify the application of some of the proposed standard *series* names to correlative *sequences* of rocks along the eastern margin of the Permian Basin from Texas to Nebraska." (Tomlinson, *et al.*, 1940, p. 342.)

"May the name of a *sequence* of sediments, first considered as a *formation*, be used later as a *series* name?" (Tomlinson, 1940, p. 356.)

"Actually the type region may contain only a part of the full *sequence* assigned in general usage to the particular *system*—for example the Devonian of Devonshire, England." (Ashley, *et al.*, 1933a, p. 446.)

Sequence also has been used with much the same general meaning when applied to faunal successions as shown by the following:

"Breaks in the faunal *sequence* at bottom and top of these systems are accompanied by more or less widespread physical breaks." (Moore, 1940, p. 290.)

"The best criterion would be the absence of a stage in the evolutional *sequence* of a lineage that is complete at a neighboring locality." (Neaverson, 1928, p. 11.)

The italics in all the above quotations are the writer's. Other examples might be cited, but these are sufficient to illustrate the point. In the writer's opinion the purpose of uniformity of usage would be served better if sequence were given the specific meaning proposed by Moore and succession used in the sense in which sequence has been employed in the above quotations.

PERIOD—SYSTEM

Because of the almost universal usage of these terms with the same meaning they need but little discussion. Like eras and sequences these divisions are world-wide in application. The periods are generally considered the fundamental units of the standard time scale, and each is defined in the same manner as an era except that the crustal deformations which delimited them were of less magnitude than for the eras, and the consequent changes in sedimentation and life less prominent. The rocks formed during a standard period of time constitute a system.

Ideally, these rocks are set off by unconformities and contain distinctive assemblages of fossils.

EPOCH—SERIES

The epoch is now generally accepted as a subdivision of a period, but that is about the only consistent point in the usage of the term. The United States Geological Survey (Wilmarth, 1925, p. 4), for example, uses the term to apply to any time interval of less duration than a period irrespective of the relative length, whether it be the time of deposition of a series, group, formation, member, or even an individual bed or layer. It seems that the use of this term for a definite time interval would do much to simplify nomenclature. It is herein advocated that it be restricted to the major subdivisions of a period. The rocks formed during an epoch of time, in this restricted sense, constitute a series, a major subdivision of a system. It is common practice to utilize the terms lower, middle, and upper to refer to the major subdivisions of a system (series) when no geographic names have been applied to these units. There also has been considerable loose usage of these terms to refer to major time divisions of epoch rank. It is advocated that these should not be used for time intervals but that instead the terms early, medial, and late should be employed when time or age is implied.

Because of the lack of world-wide synchronous diastrophic movements to delimit stratigraphic units of series rank, the geographic names applied to epochs and series must of necessity be of a provincial character. While it is true that many recognized series can be more or less closely correlated between continents, there are many which cannot be so correlated.

SUBEPOCH—SUBSERIES

These terms as herein proposed are applicable to the first subdivisions of epochs and series respectively. For most of the geologic column they are not necessary. However, it is felt that they will serve a real purpose for a few portions of geologic time and the corresponding parts of the stratigraphic column. They are essential in classifications in which Mississippian and Pennsylvanian are considered as series of the Carboniferous, if smaller divisions are to be treated in the same manner by all workers. By classing the Des Moines, Missouri, and Virgil divisions of the Pennsylvanian—when it is called a series—as subseries, it is then possible to designate their major subdivisions as groups. This is the same rank given them when the Pennsylvanian is classified as a system, and Des Moines, Missouri, and Virgil as series. The same classification is applicable to the Mississippian. For a few other portions of the column, notably the middle part of the Ordovician, these terms also serve a useful purpose.

CHRON—GROUP

Group is now in rather general usage by most workers for the major subdivisions of a series. It was originally proposed in 1881 by the International Geological Congress Commission to embrace all the rocks formed during a geological era and abandoned by the 1900 Congress. Several names are more or less commonly applied to the co-ordinate time unit. The United States Geological Survey advocates the use of epoch for this time unit and for all time intervals shorter than a period (Wilmarth, 1925, p. 4). Willard (1939, p. 131, 161, 201, 240) has used stage in a general manner to correspond to groups of rocks. Moore (1933, p. 54) has proposed that age be used for this time unit. It is believed that age possesses so much value as a general time term that to restrict it to any one definite time interval would seriously impair the usefulness of an extremely valuable general term. It is very convenient, for example, to designate certain rocks as of Paleozoic, Ordovician, or Trenton age. The criteria for delimiting chron and group are the same as for units of higher rank except that they are of less significance or magnitude than for the larger ones. Correlation between groups of different continents and often between different parts of the same continent is much less definite than for higher rank units, and, consequently, names should in all cases be more provincial than for larger units. A group should be a local or provincial subdivision of a series and contain two or more formations. The term group has been variously used with such wide application that in many cases the rocks of a given group include strata of two different series. Some workers still employ the term in that sense. The term is better applied to a subdivision of a series and, when any previously defined group is found, as a result of more detailed stratigraphic studies, to bridge the division between two series, the old name should be abandoned or redefined to apply only to a portion of one of the series involved. The use of group as herein advocated will not be approved by all. The following quotation from a recent publication (Willard, 1939, p. 379) illustrates the great divergence of opinion as to the proper use of many stratigraphic terms:

"Furthermore I protest the present use of such terms as group, formation, and member. There is a deplorable tendency to map time, not rocks. I object. *System*, I would retain in the sense of connoting the rocks of a designated period produced during a cycle. *Group* is less offensive than formation or member. It may be retained, but I prefer to see it applied to the strata laid down during a particular cyclic phase, as an *on-lap group* or *off-lap group*. Geographic names might even be tacked on in such a nomenclature."

The term chron was introduced by Williams (1901, p. 583-584) to refer to an indefinite division of geologic time. This application never met with general acceptance, and the word gradually has fallen into dis-

use. It seems to be a very appropriate term to designate this particular time interval for which we never have had a suitable, generally accepted name, and it is so proposed.

STAGE—FORMATION

The term stage used with a time unit application is a radical departure from the originally proposed meaning. The International Geological Congress in 1878 adopted it "for rock units that are subordinate to series." In America the term has come gradually to be used as a name for relatively short time intervals with special reference to subdivisions of the Pleistocene, for example, Wisconsin glacial stage, Aftonian interglacial stage. Willard (1939, p. 131, 161, 201, 240) has used stage in a general manner to embrace either an individual group or, in the case of the Chautauquan "stage," more than one group. Recently the term has been receiving wider application as a time coordinate of formation (Moore, 1933, p. 54; Schuchert and Dunbar, 1933, p. 72). The writer believes that the use of stage as a time term has become so firmly established that restriction of it to the time interval during which a geologic formation was deposited is warranted, and it is proposed that it be given this specific meaning.

The formation is the fundamental unit in local stratigraphic classification. Larger units, as groups, series, or systems, are regarded as assemblages of formations, and the smaller units, such as members and beds, are regarded as subdivisions of formations. The differentiation of formations is based on local succession of strata, divisions being made where the lithology changes or where breaks in the continuity of sedimentation occur, or where there is evidence of other significant geologic events. Formations are thus considered stratigraphic units which may represent either relatively short or long time intervals. In actual practice the formations are usually selected in such a way that they best meet the practical and scientific needs of the geologist.

The United States Geological Survey (Walcott, 1903, p. 23) made the formation the cartographic unit and defined it as follows:

> "Each formation shall contain between its upper and lower limits either rocks of uniform character or rocks more or less uniformly varied in character, as, for example, a rapid alternation of shale and limestone. When the passage from one kind of rock to another is gradual it will sometimes be necessary to depend almost entirely on the contained fossils in separating them. The selection of formations shall be such that they will best meet the practical and scientific needs of the user of the map."

It was further stated,

> "As uniform conditions of deposition were local as well as temporary, it is to be assumed that each formation is limited in horizontal extent. The formation should be recognized and should be called by the same name as far as it can be traced and

identified by means of its *lithologic character,* its *stratigraphic association,* and its *contained fossils.*" (Author's italics.)

Following these recommendations it is seen that formational units should be determined on all available evidence such as lithology, paleontology, and structural and stratigraphic relations. It is also clear that a formation should be distinguished and designated by the same name in all localities where it can be demonstrated that it is the same stratigraphic unit, that is, as far as its essential entity is maintained even though the lithology may change, either as a result of original deposition or subsequent alterations. This is somewhat at variance with the opinion recently expressed by Stockdale (1939, p. 37) who states that the same name should be used, "only insofar as it exists with the same dominant lithologic traits," and with the opinion of Willard (1939, p. 397) who

"prefers to confine the meaning of formation to a *mappable facies.* Variations of that facies such as may be produced by a temporary interruption of transgressions may be designated as members, and in so doing many actually are identical to plenty of members and formations recognized under the present system of identification."

The writer believes that the same name should be applied even if the general lithology has changed, if field relations and fossil content prove the unit to be the same. There can be no objection to referring to a formation as a limestone in one locality and a shale in another with the same formational name in both. The Paint Creek formation of western Kentucky may be cited as an example. Change from rock which is dominantly shale to one which is dominantly limestone occurs in short distances. It seems perfectly permissible to call this unit the Paint Creek limestone in one locality and the Paint Creek shale in another or simply to designate it the Paint Creek formation in both areas. This would seem to be in harmony with the policy of the United States Geological Survey as noted above.

SUBSTAGE—MEMBER

It is often advisable for scientific or practical reasons to map or recognize one or more prominent or especially developed portions of a formation. These are commonly designated as members and should be named in the same manner as formations except that when member distinction depends on some property not shown by the lithologic name alone, it is permissible to insert an adjective in the name. The interpretation of member as conceived by Willard (1939, p. 379) was given in the discussion of the formation. It is proposed that the co-ordinate time term for member shall be substage. Obviously the length of time involved in different substages varies greatly. Names of members and substages are quite provincial in application.

PHASE AND ZONE

These names were proposed by the International Geological Congress of 1900 for the fifth order subdivisions in the scheme of dual classification. They are not now commonly used with the same meanings as originally proposed. McKee (1938, p. 13) has employed the term phase in much the same way as many authors utilize lithologic facies. A zone, as commonly defined, is a rock unit characterized by a particular fossil form or a faunal or floral assemblage and should properly be designated by a name derived from a paleontological source, the name of the fossil or of some distinctive genus or species of the assemblage. A zone may be of the rank of a bed, member, formation, or even a larger unit. Zonule has been proposed (Fenton and Fenton, 1928, p. 20-22) to refer to the strata characterized by a faunule when that term is employed. This may have some value in certain cases although it has not met with general acceptance. It is herein proposed that when zone is given the meaning suggested above, phase shall be considered as the co-ordinate time term. Because zone as thus defined is not a definite subdivision of any higher ranking stratigraphic unit, the terms zone and phase were not included in the table of co-ordinate terms.

It is rather general practice, of which the writer approves, to utilize the term zone to designate lithologic units when it does not seem desirable to give to them any of the previously discussed stratigraphic ranks, as for example: "This formation contains three calcareous zones and two arenaceous zones within the general shale succession." This is the sense in which some writers at times incorrectly employ the term horizon, as will be explained. It is true that, in most instances where zone is used in this sense, conditions can be equally well expressed by the use of such terms as bed, member, or even some more general appellation as unit, if it seems more desirable to restrict zone to units characterized by certain fossils. If the term zone is applied in this manner to a lithologic unit, care should be taken to note that the usage is to connote lithology only, without any reference to the fossil or fossils which characterize it.

OTHER STRATIGRAPHIC TERMS

CYCLOTHEM

Recently it has become rather general practice for some portions of the geological column to designate alterations of nonmarine and marine strata, especially of the Pennsylvanian system, as cyclothems. Each cyclothem consists of a succession of beds, the lower of which are nonmarine and the upper marine, each general succession being set off from strata above and below by disconformities. Each cyclothem records one in-

terval of land exposure with its accumulation of sediments and vegetation followed by deposition during a marine invasion. Weller (1930, p. 99) first called these cycles. The name was later changed by him (Wanless and Weller, 1932, p. 1003) to cyclothem, which was proposed to designate

"a series of beds deposited during a single sedimentary cycle of the type that prevailed during the Pennsylvanian period. A cyclothem ranks as a formation in the scale of stratigraphic nomenclature."

Use of this term is now rather general not only in the Illinois or Eastern Interior Basin but in other parts of the country as well. With the cyclothem ranked as a formation, it seems logical to designate the various units of each cyclothem as members.

Moore (1936, p. 29) has emphasized the importance of lithologic unity in delimiting formations and has questioned the propriety of giving formational rank to cyclothems because, according to him,

"in practical use, rocks are to be classified on the basis of characters that can be observed readily in the field."

He argues that the "clearly defined lithologic entities" which are conspicuous lithologic units in the cyclothem should be considered as the ones best suited to general requirements of description, mapping, and economic use. Presumably then he would consider these of formational rank. However, he emphasizes the value of the knowledge of the cyclothems in deciphering the stratigraphy. He has proposed that cyclothems be grouped into larger units which he calls megacyclothems. However this matter may be decided, it seems that the assignment of formational rank to the cyclothems is a satisfactory solution for the Illinois Basin. If the cyclothem is considered as of formational rank, the equivalent or co-ordinate time term will be stage.

LENTIL AND TONGUE

These units belong in the same category with member, being only particular types of the latter. The term member is usually applied to a stratigraphic unit of this rank when it possesses considerable geographic extent of general blanket character or form. Lentil is more properly used if the geographic extent is not so great and the particular lithologic body possesses a more or less lenslike shape. Tongue is more properly used when the unit wedges out laterally in one general direction between sediments of different lithology.

BED, LAYER, AND STRATUM

These are the smallest units commonly recognized in stratigraphic nomenclature. They need, in general, no formal names and no coordinate time terms.

FACIES

This term is coming into more and more general usage in stratigraphic terminology. Stratigraphic facies are units of comparable age but of different lithology or of distinctive character, occurring within the same stratigraphic unit and changing laterally from one type of rock to another. The term facies should be applied to a particular lithologic development of any given stratigraphic unit irrespective of the rank of that unit if two or more different types of rock, of comparable age, compose it. It should be permissible to speak of a particular facies of a bed, member, formation, or perhaps even of a larger stratigraphic unit. It does not denote a particular stratigraphic unit which corresponds to any one definite time interval and needs no co-ordinate time term. McKee (1934, p. 13) has used phase with much the same meaning when applied to a formation and restricted facies to smaller lithologic variations within the phase despite the original proposal of the International Geological Congress of phase for a minor time unit, co-ordinate with zone as advocated above. It is believed that such usage should be discontinued because of the confusion resulting from mixing of terminology in this manner. Usage of magnafacies and parvafacies as proposed by Caster (1934, p. 19) and followed by Stockdale (1939, p. 39) may have considerable value in certain instances. A magnafacies is a lithologic facies which transgresses formational boundaries, and a parvafacies is one which is confined to a formation as generally defined. It is with the latter that most workers are concerned. It is advisable in many cases to give geographic names to facies in the same manner as to formations and members.

Facies also possesses an equally valuable application in paleontological work. Faunal or floral facies are particular fossil assemblages different from those in other occurrences of the same stratigraphic unit and of comparable age. They are associated with particular lithologic facies which reflect the conditions of sedimentation that influenced the life of that locality at the time of deposition.

HORIZON

According to the United States Geological Survey (Wood, 1916, p. 79) and the Committee on Classification and Nomenclature (Ashley, *et al.*, 1933a, p. 429) the term horizon when properly used expresses only position. A horizon has no thickness and should denote only a stratigraphic level, plane, or surface. It is often improperly used for bed or stratum or larger stratigraphic unit, as in the following: "This horizon is five feet thick" or "three thin calcareous horizons are exposed." The term is commonly used to denote the place in a section which should be occupied by a certain stratigraphic unit, present elsewhere but not rep-

resented in the given section as, "This bedding surface is the horizon of Coal No. 6." Other applications, perhaps not so accurate but nevertheless common, may be justified on the basis of long established usage. The following are cited as examples of such usage. In a given exposure of a given formation, a few feet of strata are red shale. In another near-by exposure no red shale occurs, but it is possible to determine the equivalent portion of the formation which is green shale. It is customary to express such a condition as follows: "These green shale beds represent the horizon of the red shale in the other exposure." Another example might be given. In one locality a coal seam occurs in the section, and at another exposure of the section there is no coal, but two or three smutty streaks are present. Such a condition is often referred to as follows: "The smutty streaks mark the horizon of the coal bed in the other exposure." While this manner of expressing such relations is more or less common, it is perhaps not exactly accurate because in both examples thickness is involved. It would seem that the facts or inferences in each case could be equally well expressed by other wording as: (1) "The green shale represents the same portion of the section as the red shale of the first exposure," and (2) "The smutty streaks mark the portion of the stratigraphic unit which is the coal seam in the previous exposure."

TERRANE

This term as applied to geology has been defined as follows: "a formation, or a group of formations; the area or surface over which a rock or group of rocks is prevalent." It has been used in geologic literature more or less indiscriminately with either of these meanings or with others. It has been employed to designate a large mass of the earth's crust, especially in tectonic geology. Another correct usage is to refer to the topographic surface as the "terrane" in engineering or military geology. Nace (1936, p. 2) adopted the following:

> "Terrane is employed as a term of convenience only, to designate units which include rocks belonging to both the undisputed Upper Cretaceous and Paleocene series, or which belong to both the Cretaceous and undisputed Tertiary."

It is recommended that terrane, if utilized in stratigraphic geology, should be given the meaning of an area over which a particular stratigraphic unit or several units crop out and that it not be used to designate any definite stratigraphic unit.

OTHER TIME TERMS

AGE

Age has been used by different writers to indicate time units of different magnitude. Moore (1933, p. 54) used it as a time co-ordinate for a group of rocks. The International Geological Congress proposed it

originally as a co-ordinate time term for the rock term stage, as explained earlier in this paper. The Committee on Nomenclature of the North American Continental Tertiary recently has proposed that age be used to designate a time unit which is "a fraction of an epoch", *i. e.*, the time co-ordinate of a faunal zone. This is the time interval for which phase is proposed in this paper. It is believed that age is too useful as a general time designation to be restricted to any particular time unit. It may be used in a general sense to designate the time of formation (in a historical sense) of any stratigraphic unit, as indicated by the following: rocks of Paleozoic age; of Ordovician age; of Champlainian age; or the age of any smaller unit. Other types of usage might be given.

INTERVAL AND TIME

These two terms may be employed with much the same general usage and meaning as has been suggested for age. These three terms seem to be sufficient to care for all cases which might arise for which a general time term should be needed.

HEMERA

This term was originally proposed by Buckman (Crickmay, 1933, p. 290) for an indivisible time unit. Hemeral correlation predicates only contemporaneity inasmuch as the strata in two localities were deposited, not necessarily synchronously but within the time designated by a definite hemera. Such a situation may be compared to two historic events, dated 1940, occurring not together but successively during that year and possibly several months apart. There is no stratigraphic co-ordinate of a hemera because hemerae can not be exactly delimited. This term has had practically no usage in America.

EON AND SUBERA

These terms have been used occasionally in the literature but in general fill no real need. An eon has been used to designate longer time intervals in earth history than the eras. Subera has been sparingly used to refer to a portion of an era comprised of two or more periods.

CONCLUSION

This brief paper contains no fundamentally new time or stratigraphic names. Previously defined terms are utilized with the originally proposed meanings wherever possible. Some earlier names are proposed with different applications than originally advocated where prevailing usage seems to warrant such change, and some earlier terms, now obsolete, are revived, either with the original meaning or with one somewhat different. It is believed that anything which may tend to promote more

uniform terminology in stratigraphy is worthy of consideration. Justification for such a paper as this on a perennially controversial subject is to keep alive the effort to attain ultimate agreement. Since international agreement seems least likely and local or restricted agreement most probable, it pays to keep making suggestions which may be followed by those closest home, thus creating a greater uniformity of practice. It is hoped that even if the dual classification which has been proposed is not accepted and followed in its entirety, it may be at least of some value in promoting a more consistent and uniform system of classification. It is recognized that the classification given will not meet with the approval of all geologists. Many differences of opinion have been indicated in the references to the works of others, and many more could be cited. Unquestionably these deep-seated differences will remain for many years. No universally accepted scheme of time and stratigraphic nomenclature ever has been proposed and it is almost futile to hope that one will be adopted at any time in the near future. History repeats itself, and the history of the development of stratigraphic classification and nomenclature is not likely to be an exception.

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FIGURE 1. DEEPLY DISSECTED PLATEAULIKE UPLANDS
At head of Caldwell Creek and Wood River.

FIGURE 2. CIRQUE, UPPER CALDWELL CREEK
Looking south into cirque at head of a tributary valley of Upper Caldwell Creek.

FIGURE 3. RECENT BRECCIAS, HEAD OF CALDWELL CREEK
Plastered on side of deep valley which was cut in older volcanic rocks.

FIGURE 4. CONTACT BETWEEN QUARTZ LATITE PORPHYRY (WHITE) AND VOLCANICS (DARK) AT HEAD OF CALDWELL CREEK
Volcanics at right show steep dips away from intrusion.

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STRUCTURAL AND VOLCANIC PROBLEMS IN THE SOUTHERN ABSAROKA MOUNTAINS, WYOMING

BY JOHN T. ROUSE

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ABSTRACT

The volcanic rocks of the Southern Absaroka Mountains are a southern continuation of the early and late volcanic units of the northern Absaroka Volcanic Field. Explosive and flow breccias, tuffs, and volcanic conglomerates are the dominant rock types of this region.

Throughout the greater part of the area the volcanics show essentially horizontal bedding, but in some areas steep dips are found. Along the eastern front of the

mountains, in the vicinity of Wood River, postvolcanic deformation is reflected in dips varying between 15 and 50 degrees. The general axis of the deformation strikes north-northeast, and approximately 40 square miles have been deformed.

Numerous large intrusions, representative of a fairly normal intermediate to felsic magma, occur in a belt that may be aligned in a northwest direction with intrusions in the northern Absaroka Mountains. Locally some of the intrusions have deformed the volcanics and lifted Paleozoic formations to abnormally high elevations.

Near the divide between Wood River and the headwaters of the Wind River a small area of unusual pink volcanic breccias indicates volcanic activity which may be the time equivalent of the more recent extrusions of rhyolite and basalt in Yellowstone National Park.

INTRODUCTION

LOCATION

The high precipitous mountain range, composed almost entirely of Eocene (?) volcanic rocks, on the western side of the Bighorn Basin in northwestern Wyoming is generally called the Absaroka Range. It extends northwest from the Owl Creek Mountains into Yellowstone National Park and approximately to the Wyoming-Montana state line. Fenneman (1931, p. 134) designated the southern part of this range, between the Owl Creek Mountains and the South Fork of Shoshone River, as the Shoshone Mountains. The southern section of the Absaroka Range presents problems and topography different from its northern part, but as the term "Shoshone Mountains" has not found its way into current literature the area described in this paper will be referred to as the Southern Absaroka Mountains.

PURPOSE OF STUDY

The purpose of this paper is to outline some of the more prominent structural and volcanic problems in the Southern Absaroka Mountains. Readers acquainted with this mountainous region realize that it would be physically impossible to make a complete study of the area in one field season. Further field work in the area will undoubtedly bring to light many new problems and may necessitate the revision of ideas expressed in this paper. At the same time, however, it is hoped that the material presented here may stimulate interest and serve as a guide to detailed studies.

FIELD WORK AND ACKNOWLEDGMENTS

The field work, done during the summer of 1937, was financed by a grant from the Penrose Bequest of The Geological Society of America which the writer gratefully acknowledges. The facilities of the base camp of the Yellowstone-Bighorn Research Association were at the writer's disposal and were of material aid in the field work. To members of this association, especially Professors W. T. Thom, Jr., and R. T. Chamberlin, the writer is indebted for numerous suggestions and profitable discussions concerning the field work and general problems of the

region. Mr. Sam Gould served as a very able field assistant, and thanks are due the numerous ranch owners in the area, especially Mr. Merrill Snyder, for their splendid cooperation.

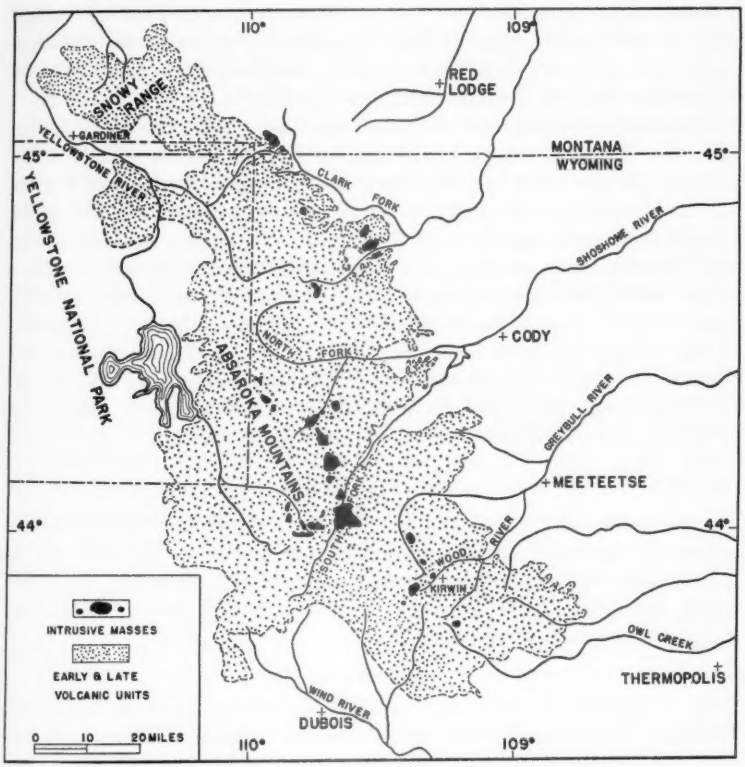

FIGURE 1.—*Sketch map of northwestern Wyoming*
Showing location of area described in this paper.

TOPOGRAPHY

The topography of the greater part of the Southern Absaroka Mountains is more rugged than that of any of the mountainous areas on the west side of the Bighorn Basin; a typical area is shown on the Kirwin quadrangle, Wyoming. The highest areas are along the divides separating the drainage of the Greybull and Wood rivers from the tributaries of the Wind River. The deeply dissected plateaulike uplands

(Pl. 1, fig. 1) at elevations of 11,000 to 12,000 feet are surmounted by high rugged peaks which may rise above 12,000 feet (Franks Peak: 13,140; Wiggins Peak: 12,160; Washakie Needles: 12,496). This plateau has been dissected by streams which have cut canyons or steep-sided valleys 3000 to 4000 feet deep. It is most inaccessible and can be reached only on foot or by horse. With considerable difficulty and extreme patience a car with high road clearance can be driven back into the mountains as far as the old mining town of Kirwin.

Forested areas are not abundant in these higher areas, the tree growth being impeded by either high elevations or exceedingly steep slopes. On the lower and less steep slopes on the east and west sides of the Wood-Wind River divide forests are more abundant, and a few small lumbering camps operate. In the main the exposures are good and not obscured by vegetation, but in some areas talus or surface wash covers critical exposures, and in other places the sheer, vertical cliffs make detailed examination of the rocks almost impossible.

The mountains have been subjected to severe glaciation, and the majority of the valley heads show well-developed cirques (Pl. 1, fig. 2). The glaciation reached maturity, and the topography is characterized by a fretted upland with abundant comb and knife ridges.

VOLCANIC ROCKS

Descriptions of the geology of the Southern Absaroka Mountains considered in this paper are few and essentially limited to Hewett's (1914, p. 121-133) brief economic report on the Kirwin District. A recent paper by Love (1939) gives an excellent detailed description of the southern margin of these mountains.

SUBDIVISIONS

Any attempt to carry the subdivisions of the northern Absaroka Volcanic Field into the Southern Absaroka Mountains meets with innumerable difficulties. In the northern Absaroka Mountains the volcanic rocks are subdivided into six units: Early Acid Breccia, Early Basic Breccia, Early Basalt Sheets, Late Acid Breccia, Late Basic Breccia, and Late Basalt Sheets. Summary description of these may be found in a previous publication (Rouse, 1937, p. 1264-1270). The early breccias may be separated from one another on the basis of mineral composition, but as the mineral compositions of the Early Acid and Late Acid as well as Early Basic and Late Basic breccias are the same it is imperative that the Early Basalt Sheets, which constitute such an excellent datum plane, be present if the units are to be placed in their correct stratigraphic position. The Early Basalt Sheets are absent in most of the

Southern Absaroka Mountains, and a subdivision into early and late units could not be made in the limited time at the writer's disposal. Love has made a detailed study of the volcanics along the southern margin of the Absaroka Mountains and tentatively correlates at least part of the Early Acid Breccia of Yellowstone Park with his Aycross formation (middle Eocene); and his Tepee Trail formation (upper Eocene) with the Late Basic Breccia (Love, 1939, p. 66-79).

It is believed that detailed mapping starting in the area worked by Love and extending north toward Yellowstone Park would show that the volcanic sequence of the Southern Absaroka Mountains could be subdivided into units which would correspond to those of the northern Absaroka Mountains.

At the head of Caldwell Creek a few isolated patches of pyroclastic rocks do not fit into the general sequence as they were extruded after the area had been dissected. These are referred to as "Recent pyroclastics".

THICKNESS

In those areas near the northeastern mountain front, where subdivisions could be made with certainty, it was found that the various units had the following thicknesses: early breccias, 800 feet ±, Early Basalt Sheets, 200 feet ±, late breccias, 200 feet ±. In the central mountainous areas where no subdivision was made the volcanics may have a total thickness of 5000 to 6000 feet.

TYPES OF PYROCLASTIC ROCKS

The volcanic sequence is represented by breccias, tuffs, volcanic conglomerates, and agglomerates. As in the northern Absaroka Mountains many of these types may be represented in a single section. Throughout this paper the classification and terminology of pyroclastic rocks suggested by Wentworth and Williams (1932, p. 45-51) is being used. The greater quantity of the pyroclastic rocks originated through explosive processes and range in composition from andesitic to basaltic. The fragments are present as sharp angular pieces which stand out in strong contrast to the tuffaceous cement. Extremely coarse breccias are uncommon in this field which is in striking contrast to the breccias of the northern Absaroka Volcanic Field. In the Southern Absaroka Mountains some of the breccias contain fragments 3 or 4 feet in diameter, but these are exceptional, and the general average size is 1 to 2 inches. Thin sections of these breccias show no alteration, fusing, or remelting of the fragments. The present field evidence indicates that these breccias are explosive products of violent volcanic eruptions. The breccias are devoid of any appreciable amount of scoriaceous or vesicu-

lar fragments. The absence of pumice or related types in the present exposures does not necessarily preclude their existence at the time of eruption. Fenner (1937, p. 236) has stated that many pyroclastics, because of their looseness and their lack of coherence, are not likely to be preserved long *in situ* unless soon covered by more resistant beds. Pumices will be carried away with considerable ease and will not be "represented in their proper relative magnitude among deposits adjacent to a volcano".

The breccias are of two general types—a darker brownish-gray in which augite is more conspicuous than hornblende, and a lighter pinkish-gray rock in which hornblende and mica predominate over augite. In addition there are numerous transitional phases between these two types making it difficult to separate the rocks into two groups, one of which will contain more silica than the other. At least 80 per cent of the breccias show well-defined bedding.

Volcanic agglomerates occur in a few isolated areas where volcanic vents may have been present. In the agglomerates the rounded fragments are much larger than the angular fragments in the breccias. A hasty trip through the Southern Absaroka Mountains might leave one with the impression that agglomerates are universally present, but more careful study will show that the rounding of the fragments in the majority of the exposures is due to stream action, and such beds cannot be called true agglomerates. The average size of the fragments in the agglomerates varies between 1 and 2 feet. The composition of the agglomerates is similar to that of the breccias.

The volcanic conglomerates are very interesting from the standpoint of origin and distribution. These are present in beds 50 to 375 feet thick, irregularly interspersed throughout the section. The matrix of the conglomerate is tuffaceous, and the majority of the pebbles, cobbles, and boulders are volcanic rocks. It is not uncommon to find boulders 1 foot in diameter in these well-bedded conglomerates, and in rare instances they may be 6 feet in diameter. These conglomerates, with their perfectly rounded pebbles and cross-bedded tuffaceous sands, undoubtedly were deposited by streams. This type of stream deposition is thought to have been similar to the type of deposition taking place in the mountains today where large alluvial fans are being deposited (Rouse, 1934, p. 750). Those familiar with this present-day type of deposition realize that at the time of heavy rains and cloudbursts the streams carry unusually large boulders, and some mud flows take place in the vicinity of the apex of the fan. The writer has seen exposures, where streams have cut through these Recent alluvial fans, in which the material is identical in structure and size with some of the Teritiary volcanic con-

glomerates. Other exposures of these volcanic conglomerates look like medium coarse river gravels. Lenses of tuffaceous sandstone, interbedded with the volcanic conglomerates, are quite common. These conglomerates are not limited to the Southern Absaroka Mountains as previous publications (Fenner, 1937, p. 237; Rouse, 1937, p. 1282) described similar types in Yellowstone Park and the northern Absaroka Mountains and called attention to the importance of streams in distributing volcanic materials in the Absaroka Mountains.

Lithic tuffs and crystal tuffs are present as well-defined beds 10 to 150 feet thick. Like the breccias they vary in composition between andesitic and basaltic.

LAVA FLOWS

At the northern end of the Southern Absaroka Mountains, on either side of the Greybull River, basalt flows appear. These are the equivalents, stratigraphically and mineralogically, of the Early Basalt Sheets of the Absaroka Volcanic Field, where they average 1000 feet in thickness. Here on the Greybull River they are only several hundred feet thick and disappear somewhere between the Greybull River and the Wood River to the south. This southward thinning and complete absence in the Wood River sections indicate that fissure eruptions of this basaltic lava did not occur in any great abundance in the Southern Absaroka Mountains.

RECENT PYROCLASTICS

The term Recent is used to differentiate the breccias emitted after much of the present topography of the Southern Absaroka Mountains had been formed. These Recent pyroclastics are much younger than the volcanic rocks just described. Several small outcrops of the Recent types are confined to the areas around the head of the Caldwell Creek (Pl. 1, fig. 3).

In this breccia the fragments of schist, gneiss, limestone, andesite, and basalt are imbedded in a pink tuffaceous matrix. The largest fragments are 2 inches in diameter, and the average size of all would be half an inch. The breccia has a uniform texture, and bedding is absent. The xenoliths of limestone and metamorphic rocks show no alteration, the limestone being as fresh and unaltered as it is in the exposures on the east side of Caldwell Creek.

In thin section the groundmass is an isotropic tuff with occasional small crystals of fresh, beautifully twinned, and unaltered andesine; equally fresh common hornblende; biotite; and a rare crystal of quartz which generally shows strain shadows. The groundmass shows no sign of flowage, and the breccia is thought to be of the explosive type.

The relation of this breccia to the older breccias is illustrated in Figure 3 of Plate 1, in which these Recent breccias are shown plastered on the side of a valley wall which had been cut by erosion prior to their extrusion. It is evident that these Recent breccias reached their present position after the Southern Absaroka Mountains, composed of the older breccias, were deeply dissected.

ORIGIN OF THE PYROCLASTIC ROCKS

GENERAL STATEMENT

The origin of the volcanic rocks in this great volcanic field is a most perplexing problem. In an earlier paper some of the problems were outlined, and suggestions were offered to help explain the origin of the pyroclastic rocks (Rouse, 1937, p. 1273-1283). In a more recent paper Parsons (1939, p. 1-26) has given an excellent description of some volcanic centers in the Sunlight Region in the northern part of the Absaroka Mountains. In the Southern Absaroka Mountains the majority of the pyroclastics have essentially horizontal bedding which leads to the conclusion that they were extruded from numerous small vents. After reaching the surface the volcanic material was distributed by one or a combination of the following processes: (a) explosion; (b) lava flows; (c) running water; and (d) mud flows.

If these vents were small then it is to be expected that extremely favorable conditions of denudation would be required to expose the centers of extrusion. Until the Southern Absaroka Mountains have been worked in detail, like the northern, the answer will remain unknown. On the basis of the present work, however, it appears that numerous vents must have been responsible for the emission of the great quantities of volcanic material. The best example of one of these small vents is Rose Butte.

ROSE BUTTE

On the west side of Rose Creek, which empties into the Greybull River above Pitchfork Ranch an old volcanic vent in the form of an isolated butte rises 500 feet above the gently sloping lower Greybull valley. The butte is not large (Fig. 2), and this may indicate that the volcanic centers in the Southern Absaroka Mountains are so small that erosion has not gone far enough to expose them.

This butte is an excellent example of an old volcanic vent in which the central portion or throat is filled with vesicular olivine basalt. Incorporated in the vesicular basalt plug are abundant fragments of petrified wood 6 to 12 inches long, 2 to 4 inches wide, and several inches thick. The central plug is surrounded by coarse, dark breccia and agglomerate in which the fragments are frequently 3 feet in diameter. The matrix

consists of soft explosive tuffaceous material and hard lava showing well-developed flow lines. The dark breccia and agglomerate are well bedded and dip steeply into the center of the vent. Mineralogically they are rich in labradorite, augite, and biotite and are basic breccias.

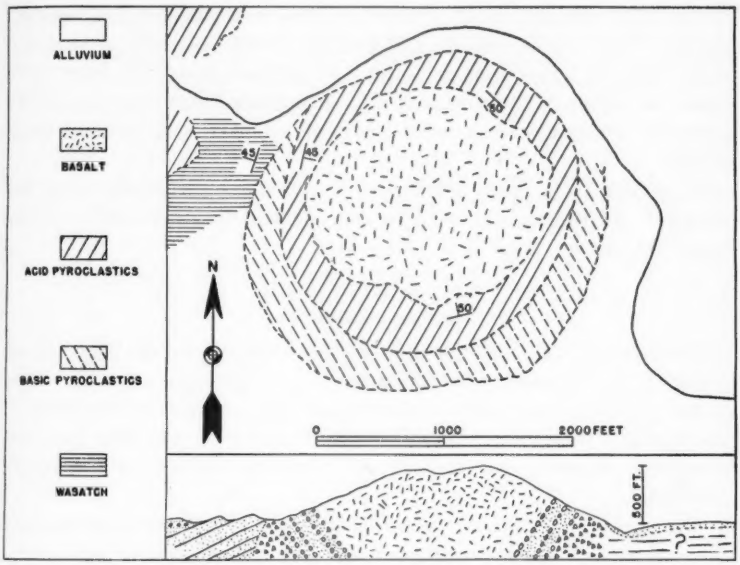

FIGURE 2.—*Geologic map and cross section of Rose Butte*

The dark basic breccias are surrounded by a band of lighter-colored acid breccias containing labradorite, hornblende, and biotite. The fragments average 5 inches in diameter and are imbedded in a pink to white tuffaceous matrix.

At the northwest end of the butte Wasatch beds dip 45 degrees away from the vent. These Wasatch beds were evidently deformed when the volcanic material was extruded from this vent.

INTRUSIVE ROCKS

Large intrusive masses are not numerous in the Southern Absaroka Mountains. These intrusions of quartz-diorite and quartz-latite-porphyry are confined to the high central part of the Mountains (Fig. 1). At the divide between Caldwell Creek and Wood River a laccolith of quartz-latite-porphyry was studied in detail (Fig. 4). The dominant

phenocrysts in this rock are plagioclase and sanidine, the former being slightly more abundant. The plagioclase is calcic oligoclase or sodic andesine and is zoned near the borders. The sanidine is zoned and present as individual small euhedral phenocrysts and as clusters of smaller crystals which are euhedral only on the outside surface of the cluster. The small amount of quartz is present as rounded and resorbed grains. The only ferromagnesian mineral found is biotite in six-sided plates (2V=0; pleochroic in shades of greenish brown). Very little biotite is visible in the thin sections. The groundmass is a fine cloudy aggregate probably composed largely of feldspar and possibly some quartz.

In addition to the larger intrusives, dikes of andesite and basalt are very common. In the main these dikes are confined to the vicinities of the larger intrusives.

STRUCTURE

LOWER WOOD RIVER

The volcanic rocks at their most eastern extension on the lower Wood River at the mountain front show abnormal structures (Fig. 3). Dips in the volcanics vary between 15 and 50 degrees, yet in exposures of the Wasatch formation, underlying the volcanics in this area, no steep dips were observed, and the Wasatch bedding planes are essentially horizontal.

The prevailing strike of the volcanics is essentially north-south, and the majority of the beds dip to the west. Beyond the western limits of the map (Fig. 3) the volcanic beds flatten rapidly. Nowhere was the transition from steeply dipping beds to horizontal beds observed. The rapidity of this change in attitude is readily apparent after a trip up the first tributary of Deer Creek. On the east side of this tributary the beds dip steeply and on the west they are practically horizontal.

In a volcanic field in which the volcanic rocks have been distributed over the surface by numerous processes—explosive, lava flows, stream and mud flows—it is to be expected that irregularities in the bedding would be observed in present exposures and that the dips observed in present-day exposures might be initial ones rather than dips produced by later deformation. The majority of the dip and strike readings shown on the map (Fig. 3) were taken on beds of volcanic conglomerate. These conglomerates have been described previously, and, if the steep dips in this area are to be explained as initial ones, then it is necessary to assume that the conglomerates were deposited on surfaces which had slopes as great as 50 degrees. Deposition of a conglomerate on such a steep slope

would be impossible, and it therefore follows that the steep dips were produced by deformation after the volcanic series had accumulated.

This deformation, reflected in the volcanic series but not in the under-

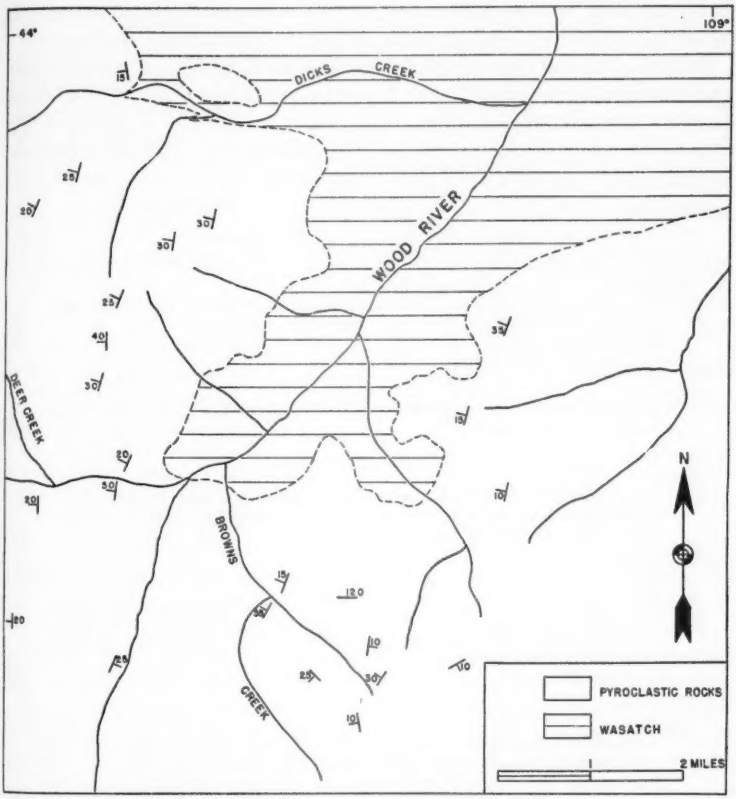

FIGURE 3.—*Geologic map of lower Wood River area*

lying essentially horizontal Wasatch beds, presents a puzzling problem in this area where there are no intrusions. On the basis of the field work done to date all that can be said is that, after accumulation of the pyroclastics in this area, rather intense but extremely local deformation took place. This deformation may have resulted from the adjustment of the volcanic rocks at the mountain front subsequent to extrusion but prior to the lithification of the volcanic rocks.

CALDWELL CREEK

An area covering approximately 4 square miles, on the divide between Caldwell Creek and the headwaters of the Wood River, shows widely divergent types of rocks and structure (Fig. 4). The interest in this area is accentuated by the abundance of good exposures. The structure is genetically related to the intrusion of quartz-latite-porphyry.

On the northwest side of the intrusion a block of Paleozoic rocks, approximately 1 square mile in plan view and 1550 feet thick, dips 30 to 45 degrees away from the intrusion. The writer is indebted to Doctor E. H. Stevens for help in measuring and correlating the uplifted Paleozoic section exposed at the head of Caldwell Creek:[1]

	Feet
TERTIARY	
Volcanic breccias	
MISSISSIPPIAN AND PENNSYLVANIAN	
Amsden	
Red shale, contains limey concretions and thin bed of limestone near base. Abundant poorly preserved fossils	30
Bedded sandstone	20
MISSISSIPPIAN	
Madison	
Interbedded dark and light limestone layers from 2-10 feet thick. Considerable lime breccia	600
DEVONIAN	
Darby	
Lime intruded by sill 100 feet thick. Brown and buff limestone and some greenish shale at top of sill. 10 feet of dark-brown to black limestone at base of sill	120
ORDOVICIAN	
Bighorn	
Dark limestone but lighter at top. Some beds fossiliferous. Beds mostly massive. Top marked by beds of limestone conglomerate and dense brown chert. Mottled zone about 50 feet above base. Round black chert nodules near base	150
Talus—no exposures	150
CAMBRIAN	
Gallatin	
Dark-gray, brown, and black limestone. Flat pebble conglomerate throughout but in thin widely separated zones	80
Limestone with some interbedded shale abundant edgewise conglomerate throughout	100
Gros Ventre	
Fine-grained brown sandstone and black shale	60
Sill	20
Thin conglomerate of rounded limestone, pebbles, and lenticular limestone	2
Black shale with a few limestone beds	50
Interbedded limestone and black shale	25
Flathead	
Interbedded quartzite and indurated shale	150
TERTIARY	
Quartz-latite-porphyry intrusion	

[1] This section has been correlated with the sections in Love's paper (1939, p. 14-29).

Insofar as is known this is the only place in the Absaroka Range where Paleozoic rocks are found in the central part of the volcanic field. Equally unique is the high elevation (maximum 12,000 feet) at which these sediments are found.

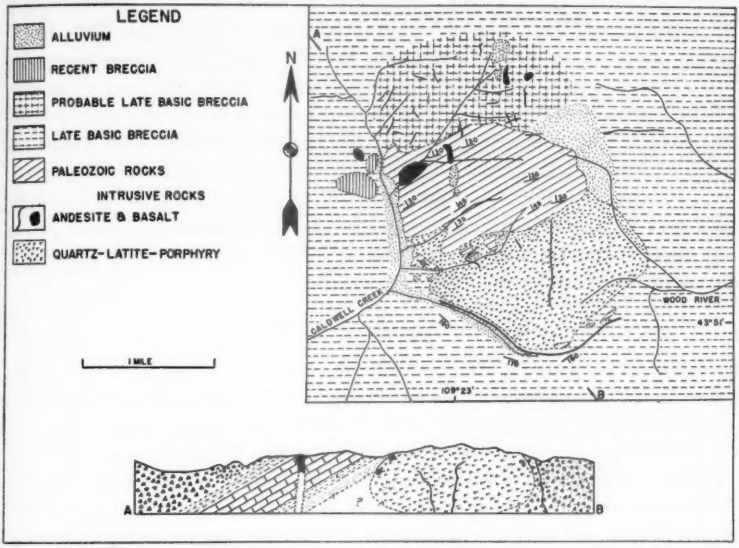

FIGURE 4.—*Geologic map and cross section of area at head of Caldwell Creek and Wood River*

With the exception of this Paleozoic block the intrusion is surrounded by volcanic rocks. Insofar as could be determined from the poor bedding, the volcanic rocks are essentially horizontal on the east, southeast, and west side of the intrusion. At the south and southwest periphery of the intrusion the volcanics dip 60 to 80 degrees to the southeast, south and southwest away from the intrusion, their strikes curving to reflect the outcrop pattern of the intrusion (Fig. 4; Pl. 1, fig. 4).

The deformation of the Paleozoic and volcanic rocks resulted from the uplifting forces accompanying the intrusion. This intrusion has many features similar to the laccolith described by Robinson (1913) at Elden Mountain in the San Francisican Volcanic Field, Arizona, where sedimentary strata have been uplifted in blocks of approximately rectangular plan and have been tilted by the intrusion of dacite.

The nearest outcrops of Paleozoic rocks are at the south and southwestern extremities of the Southern Absaroka Mountains approximately

4 miles from the intrusion and at an elevation several thousand feet below that of the uplifted block adjacent to this laccolith. The intrusion of this laccolith deformed the rocks only on the northwest and south side of the intrusion. During intrusion a block of Paleozoic rocks was tilted and uplifted on the northwest side, and at the south the volcanic beds were bent upward by the forces accompanying the intrusion. The Flathead shales show some induration, but no other metamorphic effects were observed in the uplifted Paleozoic block. On the east and west side of the intrusion the rocks show no displacement by the intrusion. Around the periphery of the laccolith numerous small limestone blocks were torn loose and metamorphosed to marble and were incorporated in the quartz-latite-porphyry. Following the intrusion of the laccolith, dikes of andesite were intruded into the quartz-latite-porphyry and surrounding country rock. A basalt dike 10 feet wide cuts the volcanics at the south and parallels the contact with the intrusion.

GENERAL STRUCTURAL PATTERN

In an earlier paper (Rouse, 1933) attention was called to a belt of intrusions trending north-northwest through the Absaroka Range and essentially parallel to the general Rocky Mountain structural trend. At that time the central area of the Southern Absaroka Mountains had not been studied. When all the intrusive bodies are placed on the map (Fig. 1) this north-northwest alignment is still more striking. These intrusions, representative of normal intermediate to felsic magmas, occupy the central higher portion of the range in areas where the volcanic rocks have their maximum thickness. The linear trend of these intrusions suggests that they were intruded along some zone of weakness which was present in the underlying basement rocks. Such a zone would also be favorable for the extrusion of volcanic material and might account for the great thickness of volcanic beds in the central area. It would appear that igneous activity might have been initiated by the extrusion of great quantities of volcanic ejectimenta through many vents located along this weak zone. After volcanic activity large quantities of magma were intruded along the same zone to form the present intrusive bodies.

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